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STATE-OF-THE-ART FOR ASSESSING EARTHQUAKE HAZARDS IN THE UNITED STATES

Report 15

TSUNAMIS, SEICHES, AND LANDSLIDE-INDUCED WATER WAVES

by

James R. Houston

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State-of-the-art methods are presented to assess the hazards of tsunamis, seiches, and landslide-induced water waves in the United States. Tsunami hazard maps for the United States are shown that display tsunami elevation zones that have a 90 percent probability of not being exceeded in a 50-year period. Methods used to determine forces exerted on structures by tsunamis are described. Hydrodynamic aspects of seiches and landslide-induced water waves are discussed, as well as methods of assessing the hazards associated with these phenomena.

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PREFACE

This report was prepared by Dr. James R. Houston of the Hydraulics Laboratory of the U. S. Army Engineer Waterways Experiment Station (WES), as part of ongoing work at WES in Civil Works Investigation Studies, "Seismic Effects of Reservoir Loading," sponsored by the Office, Chief of Engineers, U. S. Army. This is the fifteenth of a series of state-of-the-art reports for determining the hazards of earthquakes in the United States.

Preparation of the report was under the direction of Dr. E. L. Krinitzsky, Engineering Geology and Rock Mechanics Division (EG&RMD), Geotechnical Laboratory (GL). General direction was by Mr. J. P. Sale, Chief, GL, and Dr. D. C. Banks, Chief, EG&RMD.

COL Nelson P. Conover, CE, was Commander and Director of WES during the period of this study. Mr. F. R. Brown was Technical Director.

CONTENTS

CONVERSION FACTORS, U. S. CUSTOMARY TO METRIC (SI) UNITS OF MEASUREMENT

U. S. customary units of measurement used in this report can be converted to metric (SI) units as follows:

Multiply	By	To Obtain
cubic feet	0.02831685	cubic metres
cubic yards	0.7645549	cubic metres
feet	0.3048	metres
feet per second	0.3048	metres per second
feet per second squared	0.3048	metres per second squared
miles (U. S. statute)	1.609344	kilometres
miles per hour (U. S. statute)	1.609344	kilometres per hour
pounds (force)	4.448222	newtons
pounds (force) per square foot	47.88026	pascals
pounds (mass)	0.4535924	kilograms
slugs (mass)	14.5939	kilograms
slugs (mass) per cubic foot	515.3788	kilograms per cubic metre
square feet	0.09290304	square metres
tons (2000 lb, mass)	907.1847	kilograms
yards	0.9144	metres

STATE-OF-THE-ART FOR ASSESSING EARTHQUAKE HAZARDS IN THE UNITED STATES

TSUNAMIS, SEICHES, AND LANDSLIDE-INDUCED WATER WAVES

PART I: INTRODUCTION

Definitions

1. Tsunamis, seiches, and landslide-induced water waves are secondary phenomena associated with earthquakes. Tsunamis are long-period sea waves usually generated by earthquakes that cause a deformation of the sea bed. Coastal and submarine landslides and volcanic eruptions also have triggered tsunamis. The term "tsunami" originates from the Japanese words "tsu" (harbor) and "nami" (wave). This association with waves in a harbor probably is related to the large tsunami elevations that sometimes occur in harbors as a result of resonant amplification of tsunamis by partially enclosed bodies of water. Seiches are oscillations of enclosed bodies of water that may be induced by earthquakes. Oscillations produced in a harbor by a tsunami are sometimes referred to as seiches; however, a seiche is properly an oscillation of an enclosed body of water such as a lake or reservoir. Seiches are quite often produced by atmospheric phenomena rather than seismic events. Seismic events also can trigger landslides that fall into bodies of water and produce waves. These landslides may occur in coastal areas (resulting waves usually referred to as a tsunami) or enclosed bodies of water such as lakes or reservoirs.

Purpose

2. The purpose of this report is to present the state of the art for assessing the hazards of tsunamis, seiches, and landslide-induced water waves in the United States. The report is primarily

concerned with hydrodynamic aspects in assessing hazards. For example, methods to determine the hydrodynamic consequences of a landslide (given that the landslide occurred) are presented and not methods to determine the possibility that the landslide would occur.

PART II: TSUNAMIS

Background

3. Although tsunamis are secondary phenomena, they can produce great destruction and loss of life. For example, the Great Hoei Tokaido-Nanhaido tsunami of Japan killed 30,000 people in 1707. In 1868, the Great Peru tsunami caused 25,000 deaths. The Great Meiji Sanriku tsunami of 1896 killed 27,122 persons in Japan and washed away over 10,000 houses (Iida et al., 1967).

4. Tsunamis have taken many lives in the United States, with more people having died since the end of World War II as a result of tsunamis than as a result of the direct effects of earthquakes. For example, the Great Aleutian tsunami of 1946 killed 173 people in Hawaii and produced $26 million in property damage in the city of Hilo, Hawaii. The 1960 Chilean tsunami killed 61 people in Hawaii and caused $23 million in property damage (Pararas-Carayannis, 1978). The most recent major tsunami to affect the United States, the 1964 Alaskan tsunami, killed 107 people in Alaska, 4 in Oregon, and 11 in Crescent City, California, and caused over $100 million in damage on the west coast of North America (Wilson and Torum, 1968).

5. A major difference in the destructive characteristics of earthquakes and tsunamis is that earthquakes are locally destructive, whereas tsunamis are destructive locally as well as at locations distant from the area of tsunami generation. For example, the 1960 Chilean earthquake caused destruction in Chile but went unnoticed in the United States except for the recordings of seismographs. However, the tsunami generated off the coast of Chile by this earthquake not only killed over 300 people in Chile and caused widespread devastation, but also killed 61 people in Hawaii and produced widespread destruction in distant Japan where 199 people were killed, 5000 structures wrecked or washed away, and over 7500 boats wrecked or lost (Iida et al., 1967).

6. Tsunamis are principally generated by undersea earthquakes of magnitudes greater than 6.5 on the Richter scale. The typical height of

a tsunami in the deep ocean is less than a foot, and the wave period is 5 min to several hours. Tsunamis travel at [illegible] hallow-water wave celerity equal to the square root of the a[illegible]on due to gravity times the water depth even in the deepest [illegible]cause of their very long wavelengths. This speed of propagation can be in excess of 500 mph* in the deep ocean.

7. When tsunamis approach a coastal region where the water depth decreases rapidly, wave refraction, shoaling, and bay or harbor resonance may result in significantly increased wave heights. The great period and wavelength of tsunami waves preclude their dissipating energy as a breaking surf; instead, they are apt to appear as rapidly rising water levels and only occasionally as bores.

Tsunami Hazard in the United States

Atlantic and Gulf Coasts

8. The seismic activity of the Atlantic Ocean region is relatively low. In general, coasts bordering the Atlantic Ocean are not paralleled by lines of tectonic, seismic, or volcanic activity. They are rarely associated with structural discontinuities like those along the circum-Pacific seismic belt where fully 80 percent of the world's earthquakes occur. Only about 10 percent of all reported tsunamis have been in the Atlantic Ocean region.

9. The possibility of significant elevations on the Atlantic or the Gulf Coast of the United States produced by distantly generated tsunamis is very small. With the exception of the Portugal-Morocco region, the eastern Atlantic has a very low level of seismic activity. For example, the largest known shock for a thousand years in the area of Great Britain occurred in the North Sea in 1931 and had a magnitude of only 5-1/2 (Gutenberg and Richter, 1965). The Atlantic Coast of France and all of the eastern coast of Africa south of Morocco have a similar

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* A table of factors for converting U. S. customary units of measurement to metric (SI) units is presented on page 3.

low level of seismic activity. Large earthquakes do occur in certain areas of the midoceanic ridges. However, earthquakes that occur on crests of the mid-Atlantic ridge not associated with known fracture zones show either normal faulting, the tension axis being horizontal and perpendicular to the local strike of the ridge, or strike-slip motion of transform faulting. Earthquakes on the fracture zones of the mid-Atlantic ridge also are characterized by a predominance of strike-slip motion (Sykes, 1972). Large tsunamis, however, are generated by vertical ground motion (Wiegel, 1964), and only small amounts of vertical motion may accompany strike-slip motion or normal faulting with a horizontal tension axis. Consequently, although there have been many local tsunamis in the Azores Islands of the mid-Atlantic ridge, no earthquake there or anywhere along the mid-Atlantic ridge has ever produced a tsunami reported on any Atlantic coastline.

10. Large earthquakes have occurred in the Portugal-Morocco region (1356, 1531, 1597, 1722, 1755, 1761, 1773, 1926, 1960). The largest known Atlantic earthquake, and indeed one of the largest known earthquakes of historic times, occurred off the coast of Portugal on 1 November 1755. This earthquake generated the most destructive tsunami ever reported in the Atlantic. Tsunamis generated by this earthquake were reported in the West Indies. The sea rose 12 ft several times at Antigua, and every 5 min afterwards for 3 hr it rose 5 ft. The sea retired so far at St. Martin Island that a sloop riding at anchor in 15 ft of water was laid dry on her broadside. On the island of Saba, the sea rose 21 ft. At Martinique and most of the French Islands, the sea overflowed the lowland, returning quickly to its former limits (Davidson, 1936). Reid (1914), however, reported that there is little evidence that tsunamis generated by the 1755 earthquake were noticed on the coasts of the United States. The orientation of the fault along which this earthquake occurred is such that waves generated by a seismic event would be directed toward the West Indies and not the United States. Furthermore, the great continental shelf off the Atlantic and Gulf Coasts of the United States is likely to dissipate much of the energy of a tsunami. Part of the eastern coastline of Florida has a short continental shelf

and is relatively close to the West Indies. However, the shelf off the Bahamas Islands probably shelters this area.

11. In the western Atlantic, the main tsunamigenic region is the subduction zone along the arc of the West Indies Islands. The many intense earthquakes of this area have had relatively short fault lengths and, therefore, small source areas for tsunami generation. There have been no reports of tsunamis generated in this area reaching any distant coast. The only tsunami known to have been recorded on the Atlantic Coast of the United States was generated by an earthquake off the Burin Peninsula of Newfoundland on 18 November 1929. A tsunami from this Grand Banks earthquake moved up several inlets and obtained a maximum height of 50 ft. Several villages were destroyed. Tide gages on the coast of New Jersey recorded the tsunami with a 1-ft elevation at Atlantic City, New Jersey.

12. The possibility of significant locally generated tsunamis on the Atlantic or the Gulf Coast of the United States is very remote. These coastlines do not have structural discontinuities associated with seismic activity. Crustal structures have been followed by geophysical and geological methods and appear to dip far under the ocean bottom without any break (Gutenberg, 1951). Only one large earthquake has occurred on this coast in historic times. The Charleston, South Carolina, earthquake of 1886 was one of the largest earthquakes in the United States, and even one of the greatest occurring anywhere in the world. There has been no known earthquake in the Atlantic coastal plain of the United States of even remotely the same magnitude before or since (Heck, 1947). Despite the large size of the Charleston earthquake, no tsunami was generated. McKinley (1887) reported that "Except in the rivers the wave motion was not observed to have communicated to the water." Thus, this earthquake probably exhibited little of the vertical motion required to generate a significant tsunami. The complete lack of tsunamigenic activity on the eastern coast of the United States is probably a result of not only a low level of seismic activity but also the strike-slip nature of earthquakes in the area.

13. The tsunami threat from both locally and distantly generated

tsunamis is very small on the Atlantic and Gulf Coasts of the United States and, undoubtedly, less than the threat from hurricane or storm surges. However, this threat cannot necessarily be neglected when hazards are investigated for critical facilities such as nuclear power plants. For such a case, the effects must be considered of a tsunami such as that generated in Portugal in 1755, or the possibility must be investigated of the occurrence of a locally generated tsunami such as the 1929 tsunami generated off the Burin Peninsula of Newfoundland.

Puerto Rico and the Virgin Islands

14. Puerto Rico and the Virgin Islands lie along the subduction zone of the Lesser Antilles that forms the eastern boundary of the Caribbean tectonic plate. Earthquakes along this subduction zone generate important local tsunamis. Tsunamis were generated near the Virgin Islands in 1867 and 1868 (Heck, 1947). The 1867 tsunami swept the harbors of St. Thomas and St. Croix. A wall of water 20 ft high entered these harbors and broke over the lower parts of the towns. At St. Thomas, the water moved inland a distance of 250 ft. The tsunami also was large on adjacent islands and the eastern coast of Puerto Rico. The Alcalde of Yabucoa (southeastern Puerto Rico) reported that the sea retreated about 150 yd, then returned, and advanced an equal distance inland. The wave was noted as far as Fajardo (which is 20 miles to the northeast from the Alcalde of Yabucoa) and as far as 40 to 60 miles along the southern shore from the Alcalde of Yabucoa (Reid and Taber, 1919).

15. An earthquake and resulting tsunami in November 1918 killed 116 people in Puerto Rico and produced damage reported in excess of $4 million. During the tsunami, the ocean first withdrew exposing reefs and stretches of sea bottom not visible during the lowest tides. The water then returned reaching heights that were greatest near the northwest corner of Puerto Rico. At Point Borinquen, the tsunami reached an elevation of 15 ft. Near Point Agujereada, several hundred palm trees were uprooted by waves from 18 to 20 ft high. At Aguadilla, waves with heights from 8 to 11 ft were reported. The Columbus Monument, about 2-1/2 miles southwest of Aguadilla, was thrown down by waves

at least 13 ft in height, and rectangular blocks of limestone weighing over a ton were washed inland distances as great as 250 ft. Heights of 4 ft were reported at Mayaguez, and heights of 3 ft at El Boqueron. The tsunami was noticeable at Ponce, Isabela, and Arecibo, but not at San Juan. Elevations of 13 ft were reported on the west coast of Mona Island (Reid and Taber, 1919).

16. The hazard in Puerto Rico and the Virgin Islands from distantly generated tsunamis is likely to be less than the hazard from locally generated tsunamis or hurricane surges. Houston et al. (1975) demonstrated that a very large earthquake in the Portugal area similar to that of the 1755 earthquake will not produce an elevation in Puerto Rico greater than the expected elevations of locally generated tsunamis or hurricane surges.

Hawaiian Islands

17. As a result of their central location in the Pacific Ocean (where approximately 90 percent of all recorded tsunamis have occurred), the Hawaiian Islands have a history of destructive tsunamis. The earliest recorded tsunami in the Hawaiian Islands was the 1819 tsunami that was generated in Chile. Over 100 tsunamis have been recorded in the Hawaiian Islands, and 16 of these tsunamis have produced significant damage. Pararas-Carayannis (1978) compiled a detailed catalog of historical observations of tsunamis in the Hawaiian Islands.

18. The distantly generated tsunamis that have produced destruction in the Hawaiian Islands have originated from the Aleutian Islands, Chile, the Kamchatka Peninsula of the Soviet Union, and Japan. More than half of all recorded tsunamis in the Hawaiian Islands were generated in the Kuril-Kamchatka-Aleutian regions of the northern and northwestern Pacific, and one fourth were generated along the western coast of South America. Tsunamis generated in the Philippines, Indonesia, the New Hebrides, and the Tonga-Kermadec island arcs have been recorded in the Hawaiian Islands, but they have not been damaging.

19. Locally generated tsunamis also have produced destruction in the Hawaiian Islands. The 1868 tsunami that was generated on the southeastern coast of the big island of Hawaii produced severe destruction

on this coast. Runup elevations perhaps as great as 60 ft were reported during this tsunami. A tsunami generated on 29 November 1975 along the same southeastern coast of the island of Hawaii produced runup elevations as great as 45 ft. Loomis (1976) presented a detailed description of the 1975 tsunami. Cox and Morgan (1977) compiled a detailed description of locally generated tsunamis in the Hawaiian Islands.

20. The tsunami hazard in the Hawaiian Islands is not uniform. For example, elevations are generally greater on the northern side of these islands as a result of the many tsunamis generated in the Kuril-Aleutian region. Runup elevations on a single island during a tsunami also may be large at one location and small at another, even at locations that are separated by short distances. Sometimes the reasons for these variations are known. For example, the extensive reefs in Kaneohe Bay on the island of Oahu protect the bay from tsunamis by strongly reflecting or dissipating them. Often the reasons for these variations are not apparent as a result of the complex interactions that occur. Houston et al. (1977) made elevation predictions based upon historical data and numerical model calculations for the Hawaiian Islands. These predictions are discussed in Part IV.

Alaska

21. The Pacific and Americas tectonic plates collide along the subduction zone of the Aleutian-Alaskan Trench. Boundaries between tectonic plates are highly seismic with almost 99 percent of all earthquakes occurring along these boundaries (Lomnitz, 1973). The great seismicity of the region and vertical motions associated with the subduction zone make the Aleutian-Alaskan region highly tsunamigenic. The earliest recorded tsunami in this region occurred in 1788. Four major tsunamis have been generated since 1946. The 1946 tsunami was generated in the eastern Aleutian Islands, the 1957 tsunami in the central Aleutian Islands, the 1964 tsunami in the Gulf of Alaska, and the 1965 tsunami in the western Aleutian Islands.

22. Figure 1 shows a map of Alaskan localities that have experienced tsunamis. These locations are concentrated along the boundary of the Pacific and Americas plates. The remainder of Alaska has not had a

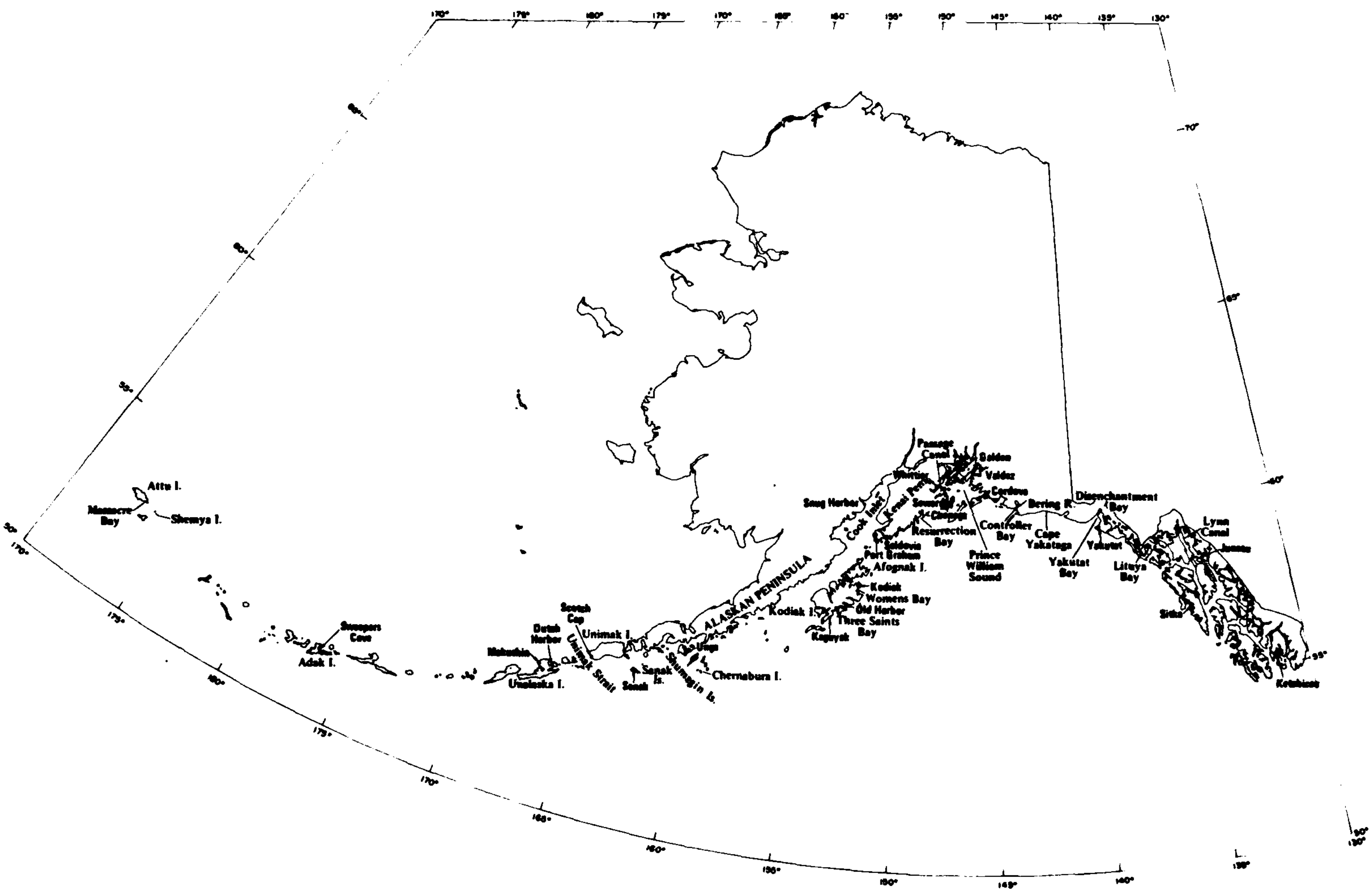


Figure 1. Map of Alaskan localities experiencing tsunamis (from Cox and Pararas-Carayannis, 1976)

reported tsunami. However, this region has a very low population density, and reporting may be quite poor. Cox and Pararas-Carayannis (1976) published a catalog of reported tsunamis in Alaska. Locally generated tsunamis dominate the catalog.

23. The 1964 Alaskan tsunami demonstrated the tremendous destructive power of major locally generated tsunamis in Alaska. This tsunami produced over $80 million in damage and killed 107 people (Wilson and Torum, 1968). In addition to the waves generated by the large-scale tectonic displacement, large waves were generated in many areas by submarine slides of thick sediments.

West coast of the
continental United States

24. The hazard on the west coast of the United States due to distantly generated tsunamis has been demonstrated by tsunami activity since the end of World War II. For example, the 1946 Aleutian tsunami produced elevations (combined tsunami and astronomical tide) as great as 15 ft above mean lower low water (mllw) at Half Moon Bay, California; 13.4 ft above mllw at Muir Beach, California; 14 ft above mllw at Arena Cove, California; and 12.4 ft above mllw at Santa Cruz, California. One person in Santa Cruz was killed by this tsunami. The 1960 Chilean tsunami produced a trough to crest height of 12 ft at Crescent City, California, and caused $30,000 in damage to the dock area and streets (Magoon, 1965). The 1964 Alaskan tsunami produced elevations above mean high water (mhw) as great as 14.9 ft at Wreck Creek, 9.7 ft at Ocean Shores, and 12.5 ft at Seaview in the state of Washington. Elevations from 10 to 15 ft above mhw were produced along much of the coast of Oregon, and four people were killed. This tsunami reached an elevation of 20.7 ft above mllw at Crescent City, California. Crescent City sustained widespread destruction with $7.5 million in damage and 11 deaths (Wilson and Torum, 1968).

25. Tsunamis generated in South America and the Aleutian-Alaskan region pose the greatest hazard (from distantly generated tsunamis) to the west coast of the United States. Historical records of tsunami occurrence in the Hawaiian Islands indicate that tsunamis generated in

the Philippines, Indonesia, the New Hebrides, and the Tonga-Kermadec island arcs do not generate tsunamis that are significant at transoceanic distances. Tsunamis, such as the 1896 Great Meiji Sanriku and the 1933 Great Shorva Sanriku that were generated off the coast of Japan, have produced no significant elevations on the west coast of the United States. Kamchatkan tsunamis, such as the ones in 1923 and 1952 (which were the greatest from Kamchatka since at least 1837), did not cause damage on the west coast. The west coast of Canada lies along a strike-slip fault that has not historically produced tsunamis on the west coast of the United States. Tsunamis off the Pacific Coast of Mexico have produced large local elevations, but they are generated by earthquakes covering areas that are apparently too small to cause significant elevations on the west coast of the United States.

26. The west coast of the United States lies along the boundary of the Pacific and Americas tectonic plates. However, this boundary is not a subduction zone. The Pacific and Americas plates have a horizontal relative motion along this boundary, and earthquakes in the region exhibit strike-slip motion that is not an efficient generator of tsunamis. For example, the 1906 Great San Francisco earthquake (8.3 magnitude on Richter scale) produced waves with heights no greater than 5 cm (Iida et al., 1967).

27. The hazard of locally generated tsunamis on the west coast of the United States is probably much less than the hazard from distantly generated tsunamis. However, there have been reports of significant locally generated tsunamis on the west coast. For example, a recent publication of the California Division of Mines and Geology (Weber and Kiessling, 1978) mentions that Wood and Heck (1966) reported that runup heights of a tsunami generated by the 1812 Santa Barbara earthquake reached 50 ft at Gaviota, 30-35 ft at Santa Barbara, and 15 ft or more at Ventura in California. However, an exhaustive study (Marine Advisors, Inc., 1965) of this event that included an investigation of the unpublished notes (cited by Wood and Heck) of the late Professor G. D. Louderback, University of California, Berkeley, has shown that the runup heights for this tsunami probably were not more than 10-12 ft at Gaviota

and correspondingly lower at the other locations. A report of a tsunami at Santa Crux, California, in 1840 also has been shown to be erroneous (Symons and Zetler, 1960). The largest authenticated locally generated tsunami on the west coast was generated by the 1927 Point Arguello earthquake and produced runup elevations as great as 6 ft in the immediate vicinity. Although there is no solid evidence that locally generated tsunamis pose a great hazard on the west coast, the possibility of significant locally generated tsunamis cannot be neglected when considering hazards to critical facilities such as nuclear power plants. There also is the possibility that locally generated tsunamis may produce greater runup elevations than are produced by distantly generated tsunamis in areas protected from distantly generated tsunamis (Puget Sound, Washington, and parts of southern California).

Pacific Ocean island territories and possessions

28. Many of the island territories and possessions of the United States are parts of seamounts that rise abruptly from the ocean floor. As a result of the very short transition distance (relative to typical tsunami wavelengths in the deep ocean) from oceanic depths to the shoreline of these islands, distantly generated tsunamis do not produce large elevations on these islands. The maximum elevation produced on such islands by distant tsunamis is on the order of 1 m (elevation recorded at Johnston Island during the 1960 Chilean tsunami by Symons and Zetler, 1960). Islands in this category include Wake Island, the Marshall Islands, Johnston Island, the Caroline Islands, the Mariana Islands, Howland Island, Baker Island, and Palmyra Island. The possibility of elevations on these islands greater than 1 m being produced by distantly generated tsunamis cannot be neglected if the hazard to critical facilities is being considered. Detailed investigations of the response of different islands to tsunamis have not been performed. It is known that 20-ft elevations were recorded on Easter Island as a result of the 1960 Chilean tsunami (generated approximately 2000 miles away). This island is small and the surrounding seamount is fairly small. The exact transition between seamounts too small to amplify

tsunamis and those large enough to cause significant amplification is not known. Numerical models discussed in Part III can be used to determine the interaction of tsunamis with islands.

29. The Samoa Islands are subject to tsunami flooding. The 1960 Chilean tsunami had a trough to crest height of 15 to 16 ft at the head of Pago Pago harbor (crest elevation of 9.5 ft) in American Samoa (Keys, 1963). Property damage of $50,000 occurred in Pago Pago village during this tsunami. Local tsunamis also are destructive in the Samoa Islands. A destructive earthquake and 40-ft tsunami have been reported to have occurred in 1917 (Heck, 1947). Whether this elevation occurred on American Samoa or one of the other Samoan Islands is not known. However, the tsunami was destructive at Pago Pago, American Samoa.

Tsunami Characteristics

Generation and deep-ocean propagation

30. Most tsunamis are generated along the subduction zones bordering the Pacific Ocean. These zones are highly seismic and earthquakes occurring within these subduction zones often exhibit the vertical dip-slip motion that is required to produce significant tsunami elevations. Berg et al. (1970) demonstrated that horizontal or strike-slip motion is a very inefficient mechanism for the generation of tsunamis.

31. Large tsunamis are associated with elliptically shaped generating areas that radiate energy preferentially in a direction perpendicular to the major axis. The major axis of the tsunami is approximately parallel to the oceanic trench or island arc that is the boundary between colliding tectonic plates. Momoi (1962) developed a relationship between the tsunami wave height H_a in the direction of the major axis of a source of length a and the wave height H_b in the direction of the minor axis of length b for an instantaneously and uniformly elevated elliptic source. This relationship is expressed by the equation $H_b/H_a = a/b$. Takahasi and Hatori (1962) demonstrated that this equation was valid by performing laboratory tests using an elliptically shaped membrane. Hatori (1963) showed that data from historical tsunamis

indicated that this equation was reasonable.

32. The directional radiation of energy from the region of generation of a tsunami is quite important. The ratio of the length of the major axis to the minor axis for large earthquakes, such as the 1964 Alaskan or the 1960 Chilean earthquake, can be approximately 4 to 6; thus the waves radiated in the direction of the minor axis can be greater than those radiated in the direction of the major axis by a similar ratio. Therefore, the orientation of a tsunami source region relative to a distant area of interest is very important, and the runup at a distant site due to the generation of a tsunami at one location along a trench cannot be considered as being representative of all possible placements of the tsunami source along the trench region. For example, the 1957 Aleutian tsunami produced significant elevations in the Hawaiian Islands but was fairly small on most of the west coast of the continental United States, whereas the 1964 Alaskan tsunami was fairly small in the Hawaiian Islands and fairly large on the northern half of the west coast. An earthquake generating a tsunami in an area southwest of the 1964 Alaskan tsunami will beam energy toward the southern half of the west coast of the United States.

33. The ground motion generating a large tsunami occurs over such a short time relative to the period of the tsunami that the motion can be considered to be instantaneous. Typical rise times (time from initiation of ground motion to attainment of permanent vertical displacement) are in tens of seconds for earthquakes, whereas tsunami periods are in tens of minutes. Higher frequency oscillations superimposed upon the movement to a permanent displacement have periods in seconds. The time for the ground rupture to move the entire length of the source is a few minutes. Hammack (1972) shows that for a large tsunami, such as the 1964 Alaskan tsunami, the actual time-displacement history of the ground motion is not important in determining far-field characteristics of the resulting tsunami. All time-displacement histories reaching the same permanent vertical ground displacement will produce the same tsunami in the far field. Hammack also showed that small-scale features of the permanent ground deformation produce waves that are not significant far

from the source region. Thus, distantly generated tsunamis can be studied knowing only major features of the permanent ground displacement.

34. Tsunamis are generated along continental margins or island arcs and then propagate out into the deep ocean. The depth transition from the relatively shallow region of generation to the deep ocean occurs over a very short distance relative to typical tsunami wavelengths in the deep ocean that are in hundreds of kilometres. In the deep ocean, tsunami wave heights are a few feet at most. The wave steepness (ratio of wave height to wavelength) is so small for tsunamis that they go unnoticed by ships in the deep ocean. Hammack and Segur (1978) demonstrated that as a result of the small amplitudes and long wavelengths of large tsunamis of consequence to distant areas, the propagation over transoceanic distances of the leading wave (or waves, since leading waves reflected off land areas may arrive at a distant location after the primary leading wave) in the generating region and in the deep ocean is governed by the linear long-wave equations. Hammack and Segur also showed that eventually nonlinear and dispersive effects will become important in the propagation of a tsunami in the deep ocean, but that the propagation distance necessary for these effects to become significant for the leading wave of a large tsunami (such as the 1964 Alaskan tsunami) is large compared with the extent of the Pacific Ocean. The later smaller waves of a tsunami wave train are shown to be ipso facto frequency dispersive (Hammack and Segur, 1978).

Nearshore effects

35. When tsunamis approach a coastal region where the water depth decreases rapidly, wave refraction, shoaling, bay or harbor resonance, and other effects may result in significantly increased wave heights. The dramatic increase in heights of tsunamis often occurs over fairly short distances. For example, during the 1960 tsunami at Hilo, Hawaii, waves could be seen breaking over the waterfront area of Hilo from a ship approximately 1 mile offshore, yet the personnel on the ship could not notice any disturbance passing by the ship. Tsunamis also can be quite large at one location and small at nearby locations (e.g.,

they may be large within a harbor as a result of resonant effects and small on the open coast).

36. As tsunamis enter shallower water, their heights increase and their wavelengths decrease; therefore, nonlinear and frequency dispersion effects become more significant. However, Hammack and Segur (1978) and Goring (1978) showed that the linear long-wave equations are adequate to describe the propagation of a large tsunami, such as the 1964 Alaskan tsunami, from the deep ocean up onto the continental shelf.

37. Tsunamis usually appear at the shoreline in the form of rapidly rising water levels but occasionally in the form of bores. When they appear as bores, frequency dispersion effects are important in the region of the face of the bore and long-wave equations are not adequate to describe flow in this region. However, beyond the face of the bore the water surface has been described as being almost flat in appearance (McDonald et al., 1947). Long-wave equations may adequately describe flows in this broad-crested region that probably governs the ultimate land inundation.

38. Even when a tsunami appears as a rapidly rising water level, there are many small-scale effects that develop that are highly nonlinear and frequently dispersive (e.g., small bores forming at the tsunami front during propagation over flatland and strong turbulence during flow past obstacles and areas of great roughness). However, there is substantial evidence that the main features of the extent of land inundation are governed by simple processes. Quite often the runup elevation (elevation of maximum inundation) is the same as the elevation near the shoreline and at other locations within the zone of inundation. Therefore, the water surface of the tsunami is fairly flat during flooding. For example, Magoon (1965) reported flooding to about the 20-ft contour above mllw and elevations at the shoreline of about 20 ft for the 1964 Alaskan tsunami at Crescent City, California. Wilson and Torum (1968) reported that the 20-ft (mllw) runup at Valdez, Alaska, for the 1964 tsunami checked "well for consistency with water-level measurements made on numerous buildings throughout the town." Similar comments were made by Brown (1964) in reference to survey measurements of 30-ft (mllw)

runup at Seward, Alaska, for the 1964 Alaskan tsunami. Runup elevations and elevations at the shoreline and in the inundation zone were similar at nine locations in Japan as recorded by Nasu (1934) in surveys following the 1933 Sanriku tsunami. This tsunami had a short period (12 min) and reached an elevation as great as 28.7 metres at one survey location. The runup elevation and the elevation near the shoreline also were similar at Hilo, Hawaii, for the 1960 tsunami (borelike waves) (Eaton et al., 1964). Differences are apparent, however, at locations where Eaton et al. demonstrated that flow divergence is significant. Flow divergence and convergence, frictional effects, and time-dependent effects (that can limit the time available for complete flooding) are probably the major effects causing differences between runup elevations and elevations near shoreline.

PART III: TSUNAMI MODELING

39. The scarcity of historical data of tsunami activity often makes it necessary to use hydraulic scale models, analytical methods, or numerical methods to model tsunamis in order to determine quantitatively the tsunami hazard. Even at locations with ample historical data, changes in land elevations and vegetation (thus changes in land roughness) as a result of development and the building of protective structures may modify the tsunami hazard. Scale models, analytical methods, or numerical methods are required to determine the magnitude of this modification.

Hydraulic Scale Models

40. Although it would not be practical to use hydraulic scale models (with reasonable scales) to model tsunami propagation across transoceanic distances, these models have found some applicaton in simulating tsunami propagation in nearshore regions and interaction with land areas. For example, hydraulic models have been used to study tsunami interaction with single islands that are realistically shaped and surrounded by variable bathymetry. Van Dorn (1970) studied tsunami interaction with Wake Island using a 1:57,000 undistorted scale model. Jordaan and Adams (1968) studied tsunami interaction with the island of Oahu in the Hawaiian Islands using a 1:20,000 undistorted scale model. They found poor agreement between historical measurements of tsunami runup and the hydraulic model data. Scale effects (e.g. viscous effects) and the effects of the arbitrary boundaries that confine the hydraulic model probably account for the poor agreement. It is also difficult to measure tsunami elevations in such small-scale models. Jordaan and Adams modeled the tsunami to a vertical scale of 1:2000; thus the waves had heights ten times the normal proportion. Even with this distortion, waves had heights of only approximately 0.3 millimetres in the model. The great expense required to build a hydraulic model of even a small island at a reasonable scale makes hydraulic models

unattractive relative to numerical models for most tsunami modeling.

41. Hydraulic models are sometimes useful in modeling complex tsunami propagation in small regions. For example, the 1960 tsunami at Hilo, Hawaii, formed a borelike wave in Hilo Bay. In addition, a phenomenon analogous to the Mach-reflection in acoustics may have developed along the cliffs north of the city (Wiegel, 1964). A Mach-stem wave may have entered Hilo and superposed upon an incident wave that came over and around the Hilo breakwater. Such complicated interactions involving borelike waves would be difficult to model numerically. However, hydraulic models have been successfully used to model the 1960 tsunami in Hilo Bay (Wiegel, 1964; Palmer et al., 1964).

42. In general, hydraulic models are not suitable for modeling tsunami propagation even within small regions. Typical tsunami wavelengths (except perhaps when borelike waves are formed) are so long that wave makers in a hydraulic model are only a small fraction of a wavelength from the region to be studied. Thus, waves reflected by land are almost immediately reflected off the wave makers and back into the basin. Wave absorber screens in front of the wave makers are not very helpful because it is very difficult to absorb very long waves. The wave makers can be moved a few wavelengths from the region of interest by reducing the model scale; however, scale effects then become very significant. Hydraulic model tests by Grace (1969) suffered the problem of trapping of wave energy between the wave makers and the shoreline.

Analytical Methods

43. Analytical solutions have long been used to study tsunamis. They are useful for simplified conditions (e.g., linear bottom slopes) rather than for general and arbitrary conditions. Analytical solutions for tsunami interactions with simple bathymetries are often used to verify numerical model solutions. For example, Omer and Hall's solution (1949) for the diffraction pattern for long-wave scattering off a circular cylinder in water of constant depth can be used to verify a numerical model that calculates the interaction of tsunamis with an

island in a constant depth ocean. Hom-ma's solution (1950) for the diffraction pattern for long-wave scattering off a circular cylinder surrounded by a parabolic bathymetry can be used to verify a numerical model that calculates the interaction of tsunamis with an island in a variable depth ocean. Analytical solutions also can provide insight into the important processes determining tsunami propagation. For example, Hammack and Segur (1978) used analytical solutions to provide criteria for the modeling of tsunami propagation.

44. Analytical solutions are often used in engineering practice to determine tsunami modification during propagation over simple bathymetric variations or interaction with simple shoreline configurations. Camfield (in preparation) presented a description of many of the analytical solutions that have been used in engineering practice. These solutions are useful when time or cost constraints rule out application of more general numerical models.

45. There are problems associated with use of analytical solutions to determine tsunami propagation for actual arbitrary conditions. Different phenomena, such as refraction, diffraction, shoaling, reflection, and runup, usually must all be solved separately if analytical solutions are used. There are many techniques used to solve each of these processes. The techniques often make different simplifying assumptions and provide different answers. For example, Camfield (in preparation) discussed several formulas that have been used to calculate tsunami runup. Some traditional techniques, such as simple refraction methods, also have been shown to be inadequate for most tsunami propagation problems (Jonsson et al., 1976).

Numerical Models

Generation and deep ocean propagation

46. Numerical models have been developed to generate tsunamis and to propagate them across the deep ocean (Hwang et al., 1972; Chen, 1973; Garcia, 1976). These models use finite difference methods to solve the

linear long-wave equations on a spherical coordinate grid. One of the models (Hwang et al., 1972) solves a nonlinear continuity equation, since the total water depth including the tsunami height is used. However, the tsunami height is so small compared with the water depth that this nonlinearity is inconsequential (and requires additional computational time). These models employ grids covering large sections of the Pacific Ocean. Transmission boundary conditions are used on open boundaries to allow waves to escape from the grid instead of reflecting back into the region of computation. Two of the models (Chen, 1972; Garcia, 1976) solve the equations of motion with an explicit formulation. The model of Hwang et al. uses an implicit-explicit formulation developed by Leendertse (1967) and the transmission boundary condition that requires the time step employed in the calculations be limited by the stability constraint for explicit formulations. However, the implicit-explicit formulation requires more computational time than required by explicit formulations when the time step is limited by the same stability constraint.

47. These generation and deep-ocean propagation models use an initial condition that an uplift of water surface in the source region is identical to the permanent vertical ground displacement produced by the tsunamigenic earthquake. Hammack (1972) demonstrated that it is this permanent vertical ground displacement, and not the transient motions that occur during the earthquake, that determines the far-field characteristics of the resulting tsunami. In addition, Hammack showed that the small-scale details of the permanent ground deformation produce waves that are not significant far from the source region. Thus, distantly generated tsunamis can be studied only knowing the major features of the permanent ground deformation.

48. Hwang et al. (1972) used data of the permanent vertical ground displacement of the 1964 Alaskan tsunami collected by Plafker (1964) in a simulation of the 1964 tsunami. Good agreement was demonstrated by Hwang et al. between a recording of the 1964 tsunami (Van Dorn, 1970) in relatively deep water off the coast of Wake Island and a simulation of this tsunami using a numerical model. Houston (1978) used the model of

Hwang et al. and the model of Garcia (1976) to generate the 1964 Alaskan and the 1960 Chilean tsunami, respectively. The data of the permanent vertical ground displacement of the 1964 and 1960 earthquakes collected by Plafker (1964) and Plafker and Savage (1970) were used as initial conditions in these models. The deepwater wave forms calculated by these models were used by Houston (1978) as input to a nearshore numerical model covering the Hawaiian Islands. Good agreement was shown between tide gage recordings of these tsunamis in the Hawaiian Islands and the numerical model calculations. Houston and Garcia (1977) showed similar comparisons between tide gage recordings of the 1964 Alaskan tsunami on the west coast of the United States and numerical model calculations.

Tsunami interaction with islands

49. Tsunami destruction in the Hawaiian Islands has directed interest toward the development of numerical models to simulate the interaction of tsunamis with islands. Several numerical models have been developed in recent years. All of these models solve the linear long-wave equations, but different techniques are used in the solutions; therefore, the models have different capabilities.

50. Vastano and Reid (1967) developed a numerical model to study the problem of determining the interaction of monochromatic plane waves of a tsunami period with a single island. A transformation of coordinates allowed a mapping of the arbitrary shoreline of an island into a circle in the image plane. The finite difference solution employed a grid that allowed greater resolution in the vicinity of the island than in the deep ocean. Such a variable grid is important since islands are usually small and surrounded by a very rapidly varying bathymetry. This numerical model can be applied only to a single island and not a multiple-island system.

51. Vastano and Bernard (1973) extended the techniques developed by Vastano and Reid (1967) to multiple-island systems. However, the transformation of coordinates technique allows high resolution only in the vicinity of one island of a multiple-island system. Thus, when Vastano and Bernard applied their model to the three-island system of

Kauai, Oahu, and Niihau in the Hawaiian Islands, the two islands of Oahu and Niihau had to be represented by cylinders with vertical walls whose cross sections were truncated wedges. Kauai was represented by a circular cylinder with the surrounding bathymetry increasing linearly in depth with distance radially from the island until a constant depth was attained. A single Gaussian-shaped plane wave composed of a broad band of wave frequencies was used as input to the model. No comparisons were made with historical tsunami data for the three islands. The model does allow the approximate effects of neighboring islands on a primary island of interest to be included in the calculations.

52. A finite difference model employing a grid covering the Hawaiian Island chain was used by Bernard and Vastano (1977) to study the interaction of a plane Gaussian pulse with the islands. The square grid cells were 5.5 km on a side and close to the minimum feasible size for a constant-cell finite difference grid covering the major islands of Hawaii. However, historical data indicate that significant variations of tsunami elevations occur over distances much less than 5.5 km. The islands of Hawaii are relatively small and not well represented by a 5.5-km grid. For example, Oahu has a diameter of only approximately 30 km, and the land-water boundary of the island has characteristic direction changes that occur over distances of much less than 5.5 km. The offshore bathymetry of the islands also varies rapidly with depth changes of more than 1500 m frequently occurring over distances of 5.5 km. Furthermore, if a resolution of eight grid cells per wavelength is maintained for tsunami periods as low as 15 min, a 5.5-km grid cannot be used for depths much below 300 m. However, the processes that cause significant modifications and rapid variations of elevations along the coastline, which are known to occur during historical tsunamis, probably occur in this region extending from water at a depth of 300 m to the shoreline. This model allows all the islands of a multiple-island system to be included in the calculations and can determine the interaction of an arbitrary tsunami with the island system.

53. Lautenbacher (1970) developed a numerical model that solved an integral equation. He applied it to an island with sloping sides

surrounded by a constant depth ocean. An advantage of the model is that a wall or "no flow" condition is not required at the shoreline. However, the computational requirements of the model are extremely large since the matrix to be inverted is full. Thus, it is not feasible to apply the model to determine the interaction of tsunamis with actual islands surrounded by complex bathymetries. In addition, Mei (1979) demonstrated that the integral equation method can have eigensolutions at certain frequencies and lead to ill-conditioned matrices.

54. A finite element numerical model based upon a model developed by Chen and Mei (1974) for harbor oscillation studies was used by Houston (1978) to calculate the interaction of tsunamis with the Hawaiian Islands. The model employed a finite element grid that telescoped from a large cell size in the deep ocean to a very small size in shallow coastal waters. The grid covered a region that included the eight major islands of the Hawaiian Islands. Although time periodic motion was assumed in the solution, the interaction of an arbitrary tsunami wave form with the islands was easily determined within the framework of a linear theory by superposition. Houston also demonstrated good agreement (major waves) between tide gage recordings of the 1960 Chilean and 1964 Alaskan tsunamis in the Hawaiian Islands and the numerical model simulations of these tsunamis. A generation and deep-ocean propagation numerical model was used to determine deep-ocean wave forms for these two tsunamis. These wave forms were used as input to the finite element numerical model. The advantages of this model include the flexibility of the finite element method that allows a telescoping grid so that extremely small elements can be placed in the nearshore region and the very small computational time required by the model as a result of the very tight bandedness of the matrix that is inverted. The model cannot be used to calculate the effects of local tsunamis generated within the Hawaiian Islands.

55. A time-stepping finite element numerical model developed by Sklarz et al. (1978) has been used to investigate locally generated tsunamis near the big island of Hawaii. Large, locally generated tsunamis occurred on the southeast side of the island of Hawaii in 1868

and 1975. Sklarz et al. attempted to simulate the 1975 tsunami by using a 5-ft uplift of the water surface in an elliptical area off the coast of the island of Hawaii as an initial condition in their numerical model. The finite element model solves the linear long-wave equations. Reasonable general agreement was demonstrated between the numerical model calculations multiplied by a factor of 4 and measurements of runup for the 1975 tsunami. Sklarz et al. also claimed that a factor of 4 will account for the difference between the infinite wall at the shoreline used in their model and the amplification of a tsunami as it moves from the shoreline to the level of ultimate runup. The advantage of this model is that the flexibility of the finite element grid allows the shape of a land mass and a complex offshore bathymetry to be well represented. It is unlikely that it would be practical to include all of the Hawaiian Islands in detail in a grid used by this numerical model because a large matrix must be inverted at each time step. Thus, the cost of running the model for a large grid would be prohibitive. However, locally generated tsunamis have historically been important only on the single island nearest the uplift generating the tsunami; thus, a grid containing a single island can be used to provide the information of practical concern for locally generated tsunamis.

Tsunami interaction with coastlines

56. Numerical models have been developed to calculate tsunami interaction with continental (or large islands such as the Japanese Islands) coastlines. Many of these models solve long-wave equations. Some of the models solve long-wave equations that include nonlinear advective and dissipative terms, and others solve the linear long-wave equations. According to Hammack and Segur (1978), Goring (1978), and Tuck (1979), the propagation of large (long-period) tsunamis (at least the initial major waves), such as the 1964 Alaskan tsunami, from the deep ocean up onto the continental shelf is governed by the linear long-wave equations. Nonlinear and frequency dispersive effects are not important during this propagation since the transition from the deep ocean to the continental shelf occurs over such a short distance that there is not sufficient time for these effects to become significant. However,

these terms may become important during propagation from the edge of the continental shelf to the shoreline. The linear long-wave equations may govern tsunami interactions with islands (neglecting effects of reefs and occasional formation of bores) as a result of the very short shelf region of small islands in the Pacific Ocean.

57. One-dimensional numerical models that solve nonlinear equations and include frequency dispersion effects have been developed. Heitner and Housner (1970) developed a one-dimensional finite element model that was used to calculate the runup of a solitary wave propagating up a linear slope. Mader (1974) used a one-dimensional model to calculate the propagation of solitary waves and sinusoidal wave trains up linear slopes and past submerged barriers. Garcia (1972) used a one-dimensional model to study short-period tsunamis that might be generated by horizontal motions of the Mendocino Escarpment off the western coast of the United States. These one-dimensional models are useful since they provide insight into the interaction of tsunamis with simple models of continental slopes. However, two-dimensional models are necessary to calculate tsunami propagation over realistic bathymetries and interaction with complex coastlines.

58. Aida (1969) developed a two-dimensional finite difference numerical model with an explicit formulation to study tsunamis generated just off the coast of Japan. Various permanent vertical ground displacements for the 1964 Niigata and the 1968 Tokachi-oki tsunamis were used as initial conditions for this model that solved the linear long-wave equations. More recently Aida (1978) applied a similar model to investigate tsunami generation and propagation for five tsunamis generated off the coast of Japan. A telescoping finite difference grid was used so that grid cells could be made smaller in selected bays where there were historical measurements of these tsunamis. The tsunami source used in the calculations was a vertical displacement of the sea bottom derived from a seismic fault model for each earthquake. A crude general agreement was shown between the numerical model calculations and the historical tide gage recordings of the simulated tsunamis. Differences between the recorded and measured tsunamis are probably largely due to inaccuracies

in the seismic fault model used to determine the vertical displacement of the sea bottom.

59. Houston and Garcia (1978) employed a two-dimensional finite difference numerical model based upon the original formulation of a tidal hydraulic numerical model by Leendertse (1967) to study tsunami interaction with the west coast of the United States. This model solves long-wave equations that include nonlinear and bottom friction terms. Leendertse's implicit-explicit multioperational method is employed in solving these equations. To verify the model, a generation and deep-ocean propagation numerical model (Hwang et al., 1972) was used to generate the 1964 Alaskan tsunami and propagate it to the west coast of the United States. The resulting wave form was used as input to this nearshore numerical model that propagated the tsunami to the shoreline. Good agreement was shown between tide gage recordings of the 1964 Alaskan tsunami at Crescent City and Avila Beach, California, and the numerical model calculations. This numerical model does not employ a telescoping grid. However, the time step used by the model is not restricted by the stability critera for an explicit model. Therefore, it is practical to use fine grid cells and a grid covering a large area since a fairly large time step can be employed.

60. A time-stepping, two-dimensional finite element numerical model has been recently developed by Kawahara et al. (1978). Unlike most finite element models that are implicit and require costly matrix inversions at each time step, this model uses a two-step explicit formulation. Thus, the computational time requirements of the model are modest and would compare favorably with explicit finite difference models. Although the finite difference formulation used by Houston and Garcia (1978) allows a much larger time step than that permitted by this model, the flexibility of the finite element grid allows a region to be covered by fewer cells (as a result of the telescoping properties of the grid) and permits the shape of coastlines to be well represented. The model solves the linear long-wave equations in deep water and the long-wave equations including nonlinear advective terms in shallower water. A simulation of the 1968 Tokachi-oki tsunami is also performed by

Kawahara et al. A crude general agreement between tide gage recordings of this tsunami and the numerical model calculations are shown (with differences probably attributable to lack of knowledge concerning the ground displacement that generated the tsunami).

61. Chen et al. (1978) developed a two-dimensional finite difference model that solves Boussinesq-type equations. These higher order equations include the effects of the nonlinear advective terms and frequency dispersion. Furthermore, Chen et al. found that the third-order term accounting for frequency dispersion produced spurious high-frequency components that caused numerical difficulties. Thus, numerical filtering had to be used to suppress these components. The model was used to simulate a nearshore tsunami off the coast of Diablo Canyon, California. Chen et al. also showed that a numerical model solving long-wave equations including nonlinear terms calculated a tsunami wave form almost identical (slightly greater amplitudes) to the wave form calculated by the model that solved the Boussinesq equations. Thus, frequency dispersion did not produce any significant effects, and Boussinesq equations were not found superior to long-wave equations. The slightly lower amplitudes calculated by the model that solved Boussinesq equations may have been caused by the numerical filtering that would tend to reduce amplitudes.

Tsunami inundation

62. The final phase of tsunami propagation involves the inundation of previously dry land. As discussed earlier, inundation patterns are often fairly simple. The tsunami appears as a rapidly rising water level, and inland flooding reaches an elevation similar to the tsunami elevation at the shoreline. However, flow divergence and convergence resulting from two-dimensional variations in topography (e.g., a narrowing canyon), frictional effects, and time-dependent effects (that can limit the time available for complete flooding) can change this simple pattern of inundation. Numerical models are required to determine inundation lines for actual tsunami propagation over complex topography.

63. Bretschneider and Wybro (1976) developed a one-dimensional model to calculate tsunami inundation. Frictional effects but not

time-dependent effects were included in the model calculations. Calculations can be performed for a series of linear ground slopes. This model is easy to apply and very economical. The main limitations of this approach are the one-dimensional and time-independent properties of the solution in addition to an assumption that the height of the tsunami decreases with the square of the velocity of the tsunami. This last assumption appears to contradict laboratory experiments performed by Cross (1966).

64. Houston and Butler (1979) described a two-dimensional and time-dependent numerical model that calculates land inundation of a tsunami. The model solves long-wave equations that include bottom friction terms. A coordinate transformation was used to allow the model to employ a smoothly varying grid that permits cells to be small in the inundation region and large in the ocean. The transformation is a piecewise reversible transformation (analogous to that used by Wanstrath, 1976) that is used independently in the x and y directions to map the variable grid into a uniform grid for the computational space. A variable grid in real space is necessary since the extent of inundation has a spatial scale much smaller than a tsunami wavelength. An implicit formulation developed by Butler (1978) is used in the finite difference model. The model was verified by simulating the 1964 Alaskan tsunami at Crescent City, California. This tsunami was very large at Crescent City (20 ft above mllw), and the Crescent City area is very complex. For example, Crescent City harbor is protected by breakwaters, some of which were overtopped and others which were not. There is a developed city area, mud flats, and an extensive riverine floodplain. Inland flooding was widespread in the floodplain area and extended as much as a mile inland. Sand dunes and elevated roads played a prominent role in limiting flooding in certain areas. Good agreement was demonstrated between historical measurements (Magoon, 1965) and numerical model calculations for high-water marks, contours of tsunami elevations above the land during propagation over previously dry land, and the extent of inundation.

PART IV: TSUNAMI ELEVATION PREDICTIONS

Predictions Based upon Historical Data

65. There are locations in the United States, such as Hilo, Hawaii, that have sufficient historical data of tsunami activity to allow reasonable tsunami elevation predictions to be made based upon the available historical data. For such locations, the historical data can be ranked from the largest to the smallest recorded elevation (largest elevation with a rank equal to 1 and the second largest elevation with a rank of 2). By dividing the rank by the total number of years of record plus one year, the frequency of occurrence of elevations equaling or exceeding a recorded elevation (mean exceedance frequency) can be defined.

66. Cox (1964) found that the logarithm of the tsunami mean exceedance frequency was linearly related to tsunami elevations for the ten largest tsunamis occurring from 1837 to 1964 in Hilo, Hawaii. Earthquake intensity and the mean exceedance frequency have been similarly related by Gutenburg and Richter (1965). Furthermore, Wiegel (1965) found the same relationship between tsunami frequency of occurrence and measured elevations for tsunamis at Hilo, Hawaii; San Francisco, California; and Crescent City, California; and Adams (1970), for tsunamis at Kahuku Point, Oahu. Rascon and Villarreal (1975) demonstrated that a linear relationship between the logarithm of the mean exceedance frequency and recorded elevations held for historical tsunamis on the west coast of Mexico (data from 1732) and on the Pacific West Coast of America, excluding Mexico. Cox showed that this linear relationship between the logarithm of the mean exceedance frequency and tsunami elevations at Hilo was valid for mean exceedance frequencies as high as approximately 0.1 per year (1-in-10-year tsunami) and that the relationship between mean exceedance frequency and elevation followed a power law for higher frequencies. Thus, the logarithmic distribution may not hold for small tsunamis. This distribution also must be invalid at some large tsunami elevation, since earthquakes reach certain maximum elevations as a result of the upper limit to the strain that can be supported by rock before

fracture (Gutenberg and Richter, 1965). Thus, tsunamis can be expected to have similar upper limits of intensity. The logarithmic distribution should be adequate (provided there is a sufficient length of historical record) to determine tsunami hazards at locations other than the sites of critical facilities such as nuclear power plants. At the site of a critical facility, the Probable Maximum Tsunami (U. S. Nuclear Regulatory Commission, 1977) must be predicted by deterministic and not probabilistic methods. Houston et al. (1975) demonstrate this type of deterministic method.

67. Other mean exceedance frequency distributions can be applied to historical data of tsunami elevations. For example, the Gumbel distribution has been used in the past to study annual stream-flow extremes (1955). Borgman and Resio (1977) illustrated the use of this distribution to determine frequency curves for nonannual events in wave climatology. If the approach of Borgman and Resio is applied to the historical data of tsunami activity in Hilo, Hawaii (as compiled by Cox, (1964)), a 1-in-100-year elevation of 28.8 ft is obtained. This compares with a 1-in-100-year elevation of 27.3 ft obtained using a logarithmic distribution. The frequency distribution governing tsunami activity at a location is not known a priori, and there is not sufficient historical data to determine a posteriori the governing distribution. However, the logarithmic distribution has been shown to provide a reasonable fit of historical data at several locations in the Pacific Ocean region.

68. Historical data of tsunami activity in the United States are available from several published sources. Iida et al. (1967) and Soloviev and Go (1969) presented catalogs of tsunami activity in the Pacific Ocean. Heck (1947) listed the worldwide tsunamis covering the period from 479 BC to 1946 AD. Beringhausen (1962, 1964) compiled a catalog of tsunamis in the Atlantic Ocean and also a separate catalog of tsunamis reported from the west coast of South America (tsunamis generated in this region are of concern to areas in the United States). Pararas-Carayannis (1978) published a catalog of tsunami activity in Hawaii, and Cox and Pararas-Carayannis (1976) a catalog of tsunami activity in Alaska. Cox and Morgan (1977) described locally generated

tsunamis in the Hawaiian Islands.

69. Detailed accounts of several major tsunamis in the United States are available. For Hawaii, Shepard et al. (1950) described the 1946 tsunami; MacDonald and Wentworth (1954), the 1952 tsunami; Fraser et al. (1959), the 1957 tsunami; Eaton et al. (1964) and USAE District, Honolulu (1962), the 1960 tsunami; and Loomis (1976), the 1975 tsunami. Wilson and Torum (1968), Brown (1964), and Berg et al. (1970) discussed the 1964 Alaskan tsunami. Magoon (1965) presented the effects of the 1960 and 1964 tsunamis in northern California. Reid and Taber (1919, 1920) discussed the 1868 tsunami in the Virgin Island and the 1918 tsunami in Puerto Rico. Keys (1963) described the 1960 tsunami in American Samoa. Symons and Zetler (1960) and Spaeth and Berkman (1967) presented tide gage recordings in the Pacific Ocean region of the 1960 and 1964 tsunamis. Tide gage records of several historical tsunamis in the Pacific Ocean are available from the World Data Center A for Solid Earth Geophysics, National Oceanographic and Atmospheric Administration, Boulder, Colorado.

Predictions Based upon Historical Data and Numerical Models

Introduction

70. Most of the coastline of the United States has little or no data of tsunami activity. For example, most of the west coast of the United States has no quantitative data of tsunami elevations. Only a very few locations have data for tsunamis other than the 1964 Alaskan tsunami. The Hawaiian Islands have substantial data of tsunami elevations for tsunamis since 1946. However, the historical observations since 1946 are at discrete locations; therefore, elevations are not known along many stretches of coastline. Data of tsunami activity since 1837 is available in the Hawaiian Islands; however, historical observations prior to 1946 are concentrated in Hilo, Hawaii.

71. In addition to the general scarcity of historical data, data that are available are for recent years when tsunami activity has apparently been greater than the long-term trend. For example, in Hilo,

Hawaii, the two largest and four of the ten largest tsunamis striking Hilo from 1837 through 1979 occurred during the 15-year period from 1946 through 1960. Two of the tsunamis from 1946 through 1960 originated in the Aleutian Islands, one in Kamchatka, and one in Chile. However, six of the ten largest tsunamis occurred during the 109-year period from 1837 through 1945 with three originating in Chile, two in Kamchatka, and one in Hawaii. Therefore, both the frequency of occurrence and place of origin of tsunamis have been remarkably variable. The exceptionally frequent occurrence of major tsunamis in Hilo, Hawaii, during the period from 1946 to 1960 is a property of the unusual activity of tsunami generation areas and not of special properties of Hilo. Thus, any analysis of tsunami activity that uses a short time span that includes the period from 1946 through 1960 will predict a significantly more frequent occurrence of large tsunamis than is warranted by historical data from 1837 through 1979.

72. From an analysis of tsunami data for Hilo, Hawaii, the errors introduced in frequency-of-occurrence calculations by consideration of a short period that includes the unrepresentative years from 1946 through 1960 are apparent. A 1-in-100-year elevation for Hilo, based upon data compiled by Cox (1964) for the ten largest tsunamis in Hilo from 1837 through 1976 and assuming a logarithmic distribution, is 27.3 ft. The 1-in-100-year elevation that is based just upon the large tsunamis during the period of accurate survey measurements in Hilo from 1946 through 1976 is 44.2 ft. Since the largest elevation in Cox's data for the 140-year period from 1837 through 1976 was 28 ft (1960 Chilean tsunami), the 44.2-ft elevation for a 1-in-100-year tsunamis is obviously much too large. The choice of frequency distributions does not change this conclusion. For example, use of a Gumbel distribution yields a 1-in-100-year elevation of 42.5 ft if the analysis is just based upon data since 1946. Of course, the quantitative accuracy of the data for tsunamis in Hilo from 1837 through 1945 may be somewhat questionable. However, there is little doubt that the recorded occurrence of large tsunamis is accurate (i.e., tsunamis noted as being significant were indeed so, and major tsunamis did not occur and go unrecorded). In

addition, errors introduced by consideration of a short period that includes the years from 1946 through 1960 are greater than the errors resulting from possible observational inaccuracies of the 19th century in Hilo. For example, increasing by 50 percent the reported elevations for the five largest tsunamis recorded in Hilo during the 19th century (these five are included in the ten largest tsunamis recorded in Hilo) yields a 1-in-100-year elevation of 30.4 ft. This elevation is similar to the 27.3-ft elevation obtained using the reported elevations for the five largest tsunamis recorded during the 19th century.

73. The lack of historical data of tsunami activity in the United States covering reasonable periods of time makes it necessary to use various methods to expand the data base. For example, Rascon and Villarreal (1975) predicted elevations at a site in Mexico by using historical data collected for the entire west coast of Mexico. A frequency distribution based upon data recorded at Hilo, Hawaii, and a Bayes estimation procedure are used to improve the estimate based upon the data for the west coast of Mexico. Such an approach is questionable since tsunamis at Hilo are primarily generated locally, in Kamchatka, in Chile, and in Alaska, whereas the tsunamis recorded on the west coast of Mexico are primarily locally generated. Therefore, there is no reason that the frequency distribution in Hilo should be related to the distribution for the west coast of Mexico. In addition, elevation predictions for the specific site in Mexico are not based upon local effects that may amplify the tsunami. The following two sections describe studies that employ various techniques, including the use of numerical models to expand the data base, and thus allow elevation predictions at arbitrary locations within the study region.

Predictions for the Hawaiian Islands

74. Houston et al. (1977) described in detail methods used to make tsunami elevation frequency of occurrence predictions for the Hawaiian Islands. In order to make these predictions it was necessary to use data of tsunami activity in Hilo, Hawaii, to expand the data base at locations having recorded data of tsunami activity since 1946. In

addition, Houston et al. used a numerical model to aid in developing predictions at locations not having complete data for tsunamis since 1946 or not having any data at all of tsunami activity.

75. To reconstruct elevations prior to 1946 at locations having historical data since 1946, Houston et al. (1977) noted that tsunamis originating near the Aleutian Islands, Kamchatka, and Chile were recorded in the Hawaiian Islands from 1946 to 1964. Therefore, the response is known of many areas in the Hawaiian Islands to tsunamis originating in the three main locations where tsunamis of destructive power in these islands have historically been generated. Houston et al. assumed that tsunamis generated in a single source region (Kamchatka or Chile, but not the Aleutians) approach the islands from approximately the same direction and have energy lying in the same band of wave periods. The difference in wave elevations at the shoreline in the Hawaiian Islands produced by tsunamis generated at different times in the same region was attributed mainly to differences in deepwater wave amplitudes. For example, the 1841 tsunami from Kamchatka produced a wave elevation in Hilo, Hawaii, that was approximately 25 percent greater than that of the 1952 tsunami from Kamchatka. The same relative magnitudes of the two tsunamis were used for all of the islands to determine the elevation that must have occurred in 1841 at some location, knowing the elevation that did occur in 1952. Therefore, knowing the elevations of tsunamis from 1946 to 1960 at a location and the response of Hilo to tsunamis from 1837 to 1960 allowed a reconstruction of the elevations that occurred prior to 1946 at the location but were not recorded (for tsunamis from Chile and Kamchatka). Data from 1837 at Hilo were used instead of data from 1837 at Honolulu (Hilo and Honolulu are the only two locations with substantial data since 1837) since data do not exist at Honolulu for the 1868 and 1877 tsunamis, and the 1837 and 1841 elevations given by Pararas-Carayannis (1978) were drops in the water level and not runup elevations.

76. The assumption that tsunamis generated in Kamchatka and Chile approach the Hawaiian Islands from nearly the same direction was justified by Houston et al. (1977) by the small spatial extent of the known

generation areas in Kamchatka and a study of tsunami propagation from Chile by Garcia (1976) that indicated that directional effects for tsunamis originating along the Chilean coast are small in the Hawaiian Islands (probably because the generation areas in Chile subtend a relatively small angle with respect to the Hawaiian Islands). The position of the Aleutian-Alaskan Trench relative to the Hawaiian Islands does introduce important directional efforts for tsunamis generated in the Aleutian-Alaskan area. However, these effects are known from historical observations for tsunamis generated in the western Aleutians (1957), central Aleutians (1946), and eastern Alaskan area (1964).

77. Historical observations of tsunamis in Hawaii support the approach by Houston et al. (1977) that estimates the elevations produced by tsunamis from Chile or Kamchatka prior to 1946 based upon data for tsunamis from these tsunamigenic regions recorded during the years of accurate survey measurements since 1946. Eaton et al. (1964) noted that in the Hawaiian Islands "Tsunamis of diverse geographic origin are strikingly different, whereas those from nearly the same origin are remarkably similar." Wybro (in preparation) showed that even the distributions of normalized elevations (elevations normalized by the largest recorded elevation) produced in the Hawaiian Islands by different Aleutian-Alaskan tsunamis are nearly the same yet quite different from the distributions for tsunamis of other origins. Therefore, it is a reasonable assumption that tsunamis from the same geographic origin produce similar runup patterns in the Hawaiian Islands. Thus, the elevation of a pre-1946 tsunami at a location that has a recorded elevation for a post-1946 tsunami from the same geographic origin can be estimated using the ratio of recorded elevations of both tsunamis at Hilo, Hawaii.

78. There are many locations in the Hawaiian Islands that do not have recordings of tsunami elevations since 1946 or only have recordings of some of these tsunamis. To reconstruct elevations at these locations, Houston et al. (1977) used a finite element numerical model covering all of the Hawaiian Islands to simulate tsunami interactions with these islands. The numerical model calculations were then used to interpolate between recorded elevations and predict elevations at locations lacking

historical observations. The finite element model and the verification simulations of actual historical tsunamis in the Hawaiian Islands are described by Houston (1978). The numerical model calculations allow predictions of the elevations of tsunamis since 1946 to be made at any location in the Hawaiian Islands. The historical record at Hilo, Hawaii, can then be used to reconstruct elevations for tsunamis prior to 1946. Thus, a record of tsunami activity dating back to 1837 (beginning of Hilo record) can be reconstructed at any location and frequency of occurrence curves determined. Houston et al. (1977) presented frequency of occurrence curves for all of the coastline of the Hawaiian Islands.

Predictions for the west coast of the United States

79. Unlike the Hawaiian Islands, the west coast of the continental United States lacks sufficient data to allow tsunami elevation predictions to be made based upon local historical records of tsunami activity. Virtually all of the west coast is completely without data of tsunami occurrence, even for the prominent tsunami of 1964. Only a handful of locations have historical data for tsunamis other than the 1964 tsunami.

80. The lack of historical data of tsunami activity on the west coast of the United States necessitates the use of numerical models to predict runup elevations. Brandsma et al. (1977) used a deep-ocean numerical model to predict probable maximum tsunami wave forms in water depths of 600 ft off the west coast of the United States. Houston and Garcia (1974) used a numerical model to predict tsunami elevations in the southern California region; Garcia and Houston (1975), numerical models to predict tsunami elevations in Puget Sound, San Francisco Bay, and Monterey Bay; and Houston and Garcia (1978), numerical models to predict tsunami elevations on all of the west coast of the United States outside of these regions.

81. In order to predict tsunami elevations on the west coast of the United States, it is necessary to base the analysis on historical data of tsunami generation in the tsunamigenic regions of the Pacific

Ocean of concern to the west coast. Houston and Garcia (1974) showed that the Aleutian-Alaskan area and the west coast of South America are the tsunamigenic regions of concern to the west coast. These regions have sufficient data on the generation of major tsunamis to allow a statistical investigation of tsunami generation. It is necessary to use historical data of tsunami occurrence in generation regions to determine occurrence probabilities of tsunamis rather than using data of earthquake occurrence to predict tsunami occurrence. Earthquake occurrence statistics are of little value since a satisfactory correlation between earthquake magnitude and tsunami intensity has never been demonstrated. Not all large earthquakes occurring in the ocean even generate noticeable tsunamis. Furthermore, earthquake parameters of importance to tsunami generation, such as focal depth, rise time, and vertical ground motion, have only been measured for earthquakes occurring in recent years.

82. Houston and Garcia (1978) used the most recent and complete catalog (Soloviev and Go, 1969) of tsunami occurrence in the Pacific Ocean to determine relationships between tsunami intensity and frequency of occurrence for the Aleutian-Alaskan and South American regions. The tsunami intensity scale used in the analysis is a modification by Soloviev and Go of the standard Imamura-Iida tsunami intensity. Intensity is defined as

$$i = \log_2(\sqrt{2}\, H_{avg}) \qquad (1)$$

This definition in terms of an average runup H_{avg} (in metres) over a coast instead of a maximum runup elevation at a single location (used for the standard Imamura-Iida scale) tends to eliminate any spurious intensity magnitudes caused by often observed anomalous responses (due, for example, to local resonances) of single isolated locations. Houston and Garcia (1978) assumed that the logarithm of the tsunami frequency of occurrence was linearly related to the tsunami intensity and used linear regression of the historical data to determine the probability distributions of tsunami generation for these two tsunamigenic regions.

83. To relate the probability distributions of different intensity tsunamis to source characteristics, Houston and Garcia (1978) assumed that the ratio of the source uplift heights producing two tsunamis of different intensities (as defined earlier) was equal to the ratio of the average runup heights produced on the coasts near these tsunami sources. This ratio is equal to $2^{(i_1-i_2)}$ for two tsunamis with intensities i_1 and i_2 .

84. The directional radiation of energy from tsunami source regions was described in Part II. The strong directional radiation from large tsunami sources makes the orientation of a tsunami source relative to a distant site where runup is to be determined very important. Thus, the runup at a distant site due to the generation of a tsunami at one location along a trench cannot be considered as being representative of all possible placements of the tsunami source in the entire trench region. In order to account for the effects of directional radiation, Houston and Garcia (1978) segmented the Aleutian and Peru-Chile Trenches and used a deep-ocean propagation model to generate tsunamis in each of the segments. The Aleutian Trench was segmented into 12 sections and the Peru-Chile Trench into 3 sections. The Aleutian Trench was segmented much finer than the Peru-Chile Trench since the Aleutian Trench is oriented relative to the west coast such that elevations produced on the west coast are very sensitive to the exact location of a source along the Trench. For example, the 1946 and 1957 Aleutian tsunamis did not produce large elevations on the west coast, whereas the 1964 Alaskan tsunami radiated waves toward the northern part of this coast where large elevations were recorded. Uplifts along the Peru-Chile Trench do not radiate energy directly toward the west coast regardless of their position along the Trench. The Peru and Chile sections of the Peru-Chile Trench also have constant orientations relative to the west coast of the United States; therefore, elevations on the west coast of the United States are relatively insensitive to source location within these sections.

85. Houston and Garcia (1978) used deep-ocean propagation numerical models to generate tsunamis with intensities from 2 to 5 in steps of

one-half intensity in each of the segments of the two trench regions. Tsunamis with intensities less than 2 are too small to produce significant runup on the west coast. An upper limit of 5 was chosen because the greatest tsunami intensity ever reported was less than 5 (Soloviev and Go, 1969). Perkins (1972) and McGarr (1976) demonstrated that future earthquakes cannot have seismic moments (measure of earthquake magnitude for large earthquakes) much larger than those of earthquakes that have occurred in recorded history. Since earthquakes only reach certain maximum magnitudes, tsunamis can be expected to have similar upper limits to intensity. The tsunamis generated in the trench regions were propagated across the deep ocean using the deep-ocean propagation models.

86. As tsunamis approach the west coast of the United States their wavelengths decrease as a result of the decreasing water depths. The numerical grids used by Houston and Garcia (1978) for deep-ocean propagation have too large a grid cell spacing to properly simulate tsunami propagation over the continental shelf of the west coast. Houston and Garcia (1974) used an analytic solution to propagate tsunamis over the continental shelf to the shoreline. Houston and Garcia (1978) used a numerical model that solved long-wave equations, including nonlinear and dissipative terms, and employed a very fine grid to propagate tsunamis over the continental shelf to the shoreline. Wave forms propagated to the west coast by the deep-ocean propagation models were the input to this nearshore numerical model. Each wave form was propagated from a water depth of 500 metres to shore using the nearshore model. Numerical simulations of the 1964 tsunami at Crescent City and Avila Beach, California, were used to verify the numerical model. At each numerical grid location on the west coast, a group of 105 wave forms were determined by Houston and Garcia (1978)--seven wave forms (for intensities from 2 to 5 in one-half intensity increments) for each segment of the Aleutian and Peru-Chile Trenches. Each of these wave forms had an associated probability equal to the probability that a certain intensity tsunami would be generated in a particular segment of a trench region.

87. The maximum "still-water" elevation produced during tsunami

activity is the result of a superposition of tsunamis and tides. Therefore, the statistical effect of the astronomical tides on total tsunami runup must be included in a predictive scheme. Houston and Garcia (1974) used an analytical solution to determine combined tsunami and astronomical tide cumulative probability distributions. Houston and Garcia (1978) also employed a direct numerical solution similar to that used by Petrauskas and Borgman (1971) to determine combined tsunami and astronomical tide cumulative probability distributions. It was necessary to employ a numerical solution since the tsunami wave forms calculated by Houston and Garcia (1978) using a nearshore numerical model did not have a simple form (e.g., sinusoidal). Houston et al. (1977) did not need to consider the effect of the astronomical tides in their elevation predictions for the Hawaiian Islands since the tidal range is quite small for these islands and the local historical data implicitly contained the effects of the astronomical tides.

88. In order to perform a convolution of tsunamis and astronomical tides, Houston and Garcia (1978) calculated tidal elevations for a year at locations all along the west coast using harmonic analysis methods (Shureman, 1941). The year was then divided into 15-min segments, and 24-hr tsunami wave forms were allowed to arrive at the beginning of each of these 15-min segments and then superposed upon the astronomical tide for the 24-hr period. The maximum combined tsunami and astronomical tide elevation over the 24-hr period was determined for tsunamis arriving at each of these 15-min starting times during a year. All of the maximum elevations had an associated probability equal to the probability that a certain intensity tsunami would be generated in a particular segment of the two trench regions and arrive during a particular 15-min period of a year. These many maximum elevations with associated probabilities were used by Houston and Garcia (1978) to determine cumulative probability distributions of combined tsunami and astronomical tide elevations. The 100- and 500-year elevations were determined for locations along the west coast of the United States using these cumulative probability distributions. Elevations for arbitrary return periods can be obtained by assuming that the 100- and 500-year elevations determined by Houston

and Garcia (1978) follow a logarithmic distribution.

Risk Calculation

89. The average frequency of occurrence F calculated by Houston et al. (1977) and Houston and Garcia (1978) is a mean exceedance frequency, i.e., an average frequency per year of tsunamis occurring and producing an equal or greater elevation. It also is possible to calculate the chance of a given elevation being exceeded during a particular period of time. Such a calculation is a risk calculation.

90. Tsunamis are usually caused by earthquakes, and earthquakes are often idealized as a generalized Poisson process (Newmark and Rosenblueth, 1971). Many investigators have assumed that tsunamis also follow a stochastic process (Wiegel, 1965; Rascon and Villarreal, 1975). The probability that a tsunami with an average frequency of occurrence of F is exceeded in D years, assuming that tsunamis follow a Poisson process, is given by the following equation:

$$P = 1 - e^{-FD} \tag{2}$$

For example, the probability that a 1-in-100-year elevation will occur in a 50-year period is

$$\begin{aligned} P &= 1 - e^{-(0.01)(50)} \\ &= 1 - e^{-0.05} \\ &= 1 - 0.61 \\ &= 0.39 \end{aligned}$$

Tsunami Hazard Maps

91. Figure 2 is the general tsunami hazard map for the United States, and Figures 3 through 11 are the detailed maps (Houston et al., 1977; Houston and Garcia, 1974 and 1978; Garcia and Houston, 1975). Houston et al. (1977) presented frequency curves of tsunami elevations

for the Hawaiian Islands, and Houston and Garcia (1974 and 1978) and Garcia and Houston (1975) predicted 100- and 500-year elevations for the west coast of the continental United States. A tsunami elevation with a 90 percent probability of not being exceeded in 50 years represents a 475-year elevation. This is easily calculated from the previous paragraph by setting $D = 50$ and $P = 0.1$ (10 percent probability of being exceeded) and solving for $1/F$.

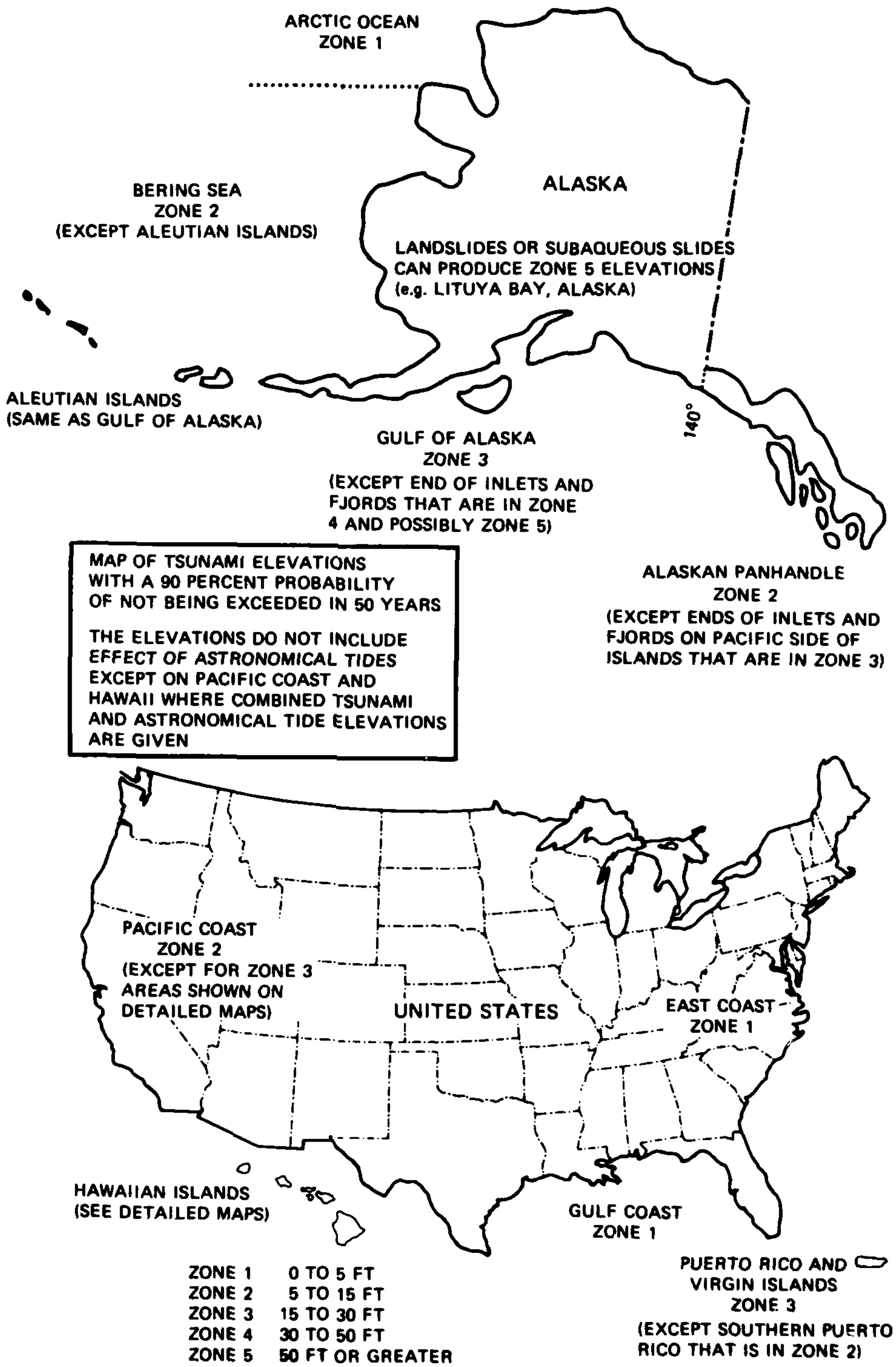


Figure 2. Tsunami hazard map

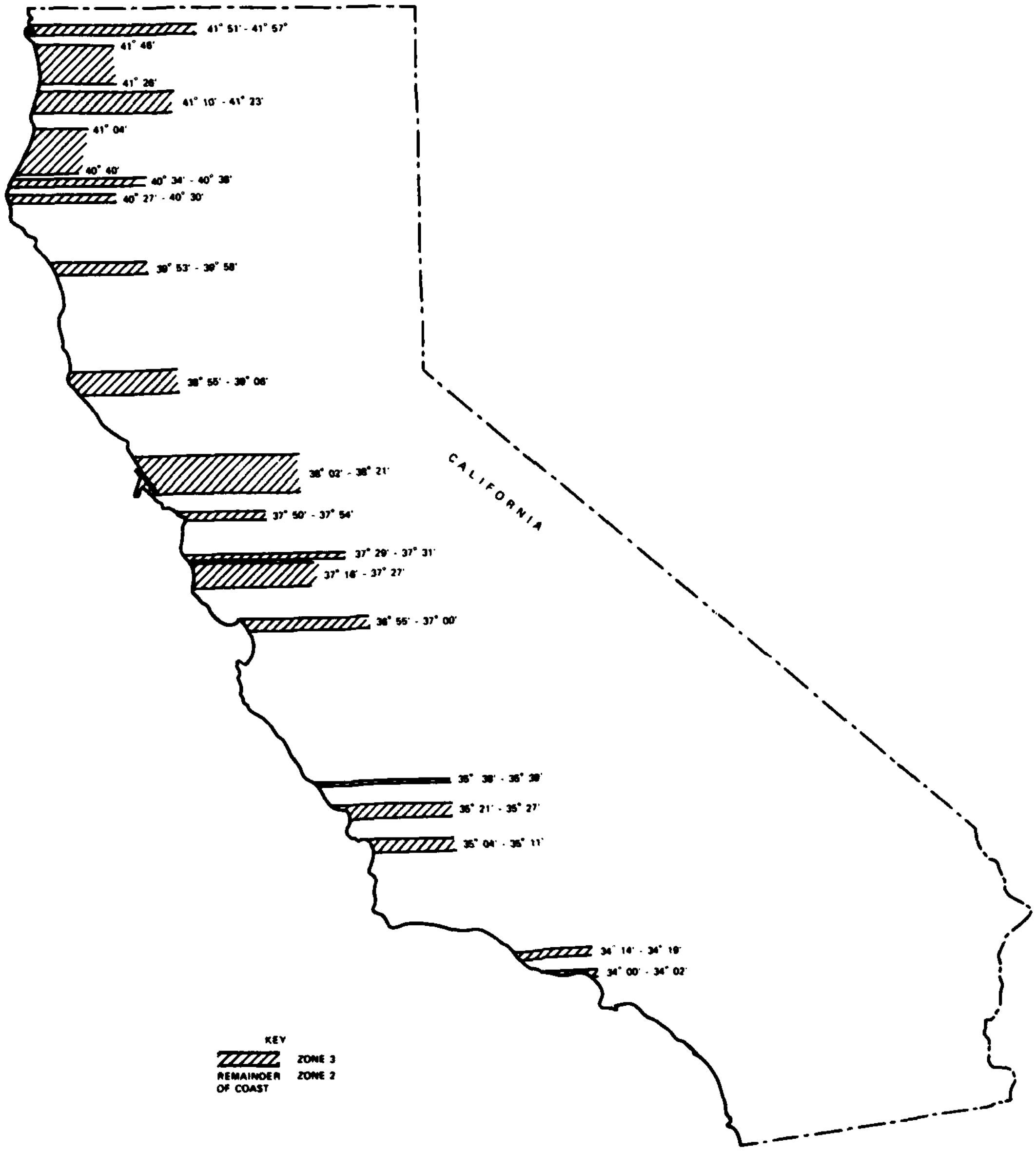


Figure 3. Tsunami hazard for California (adapted from Houston and Garcia, 1974 and 1978; Garcia and Houston, 1975)

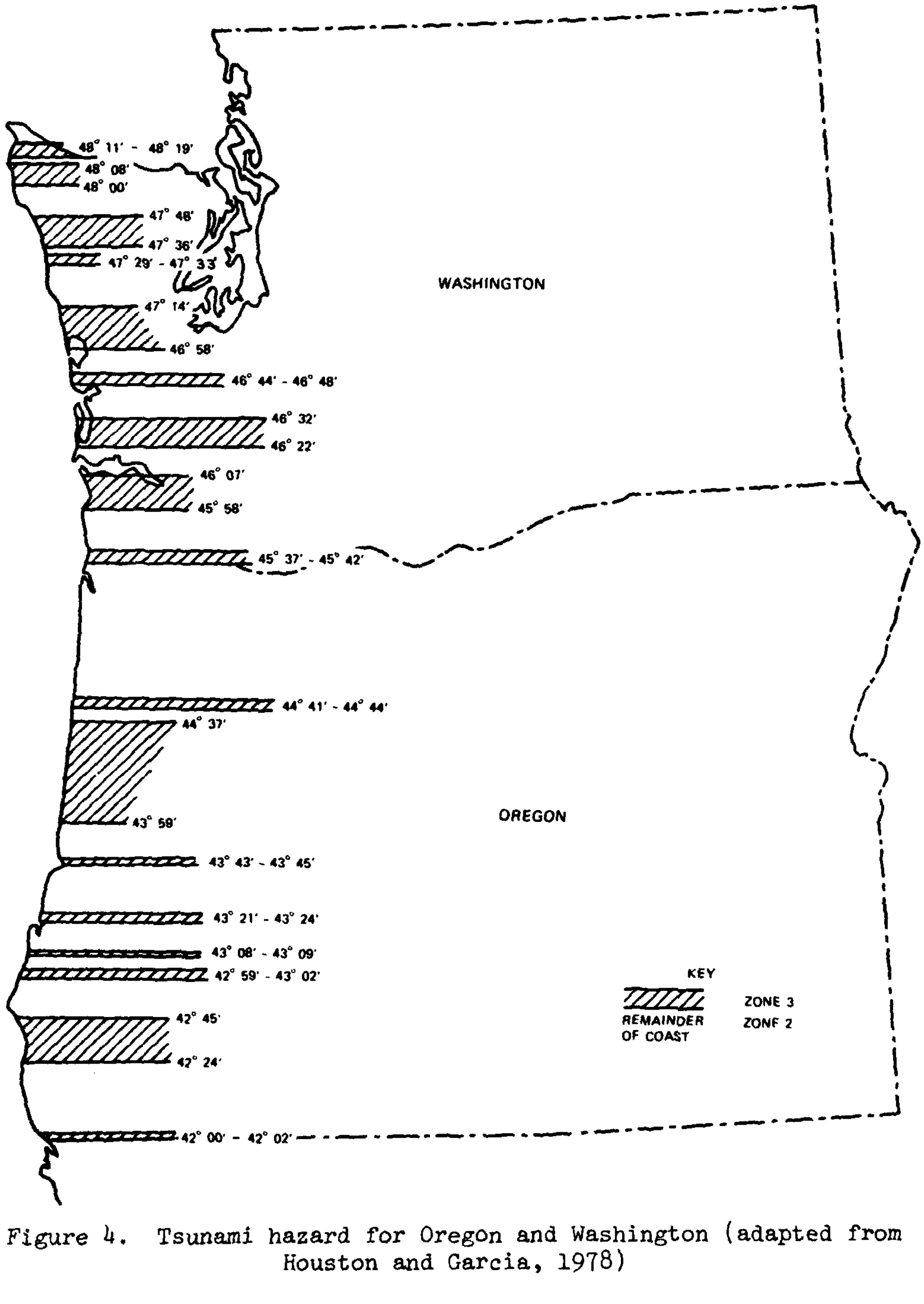
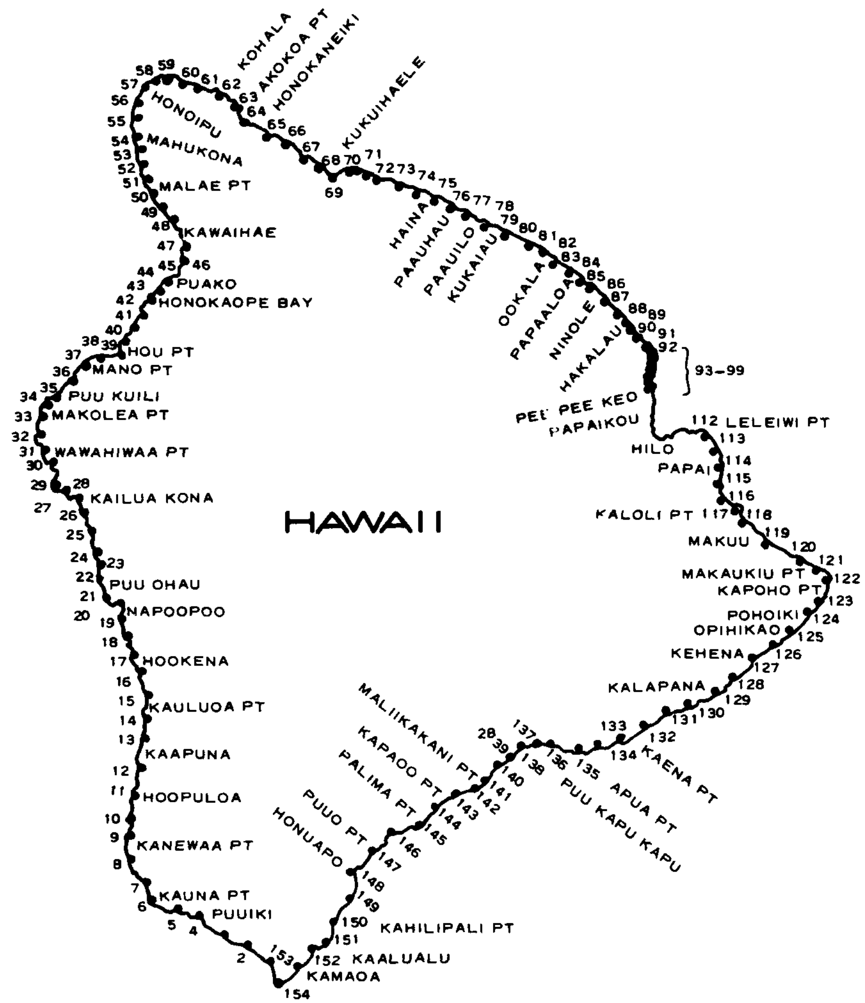
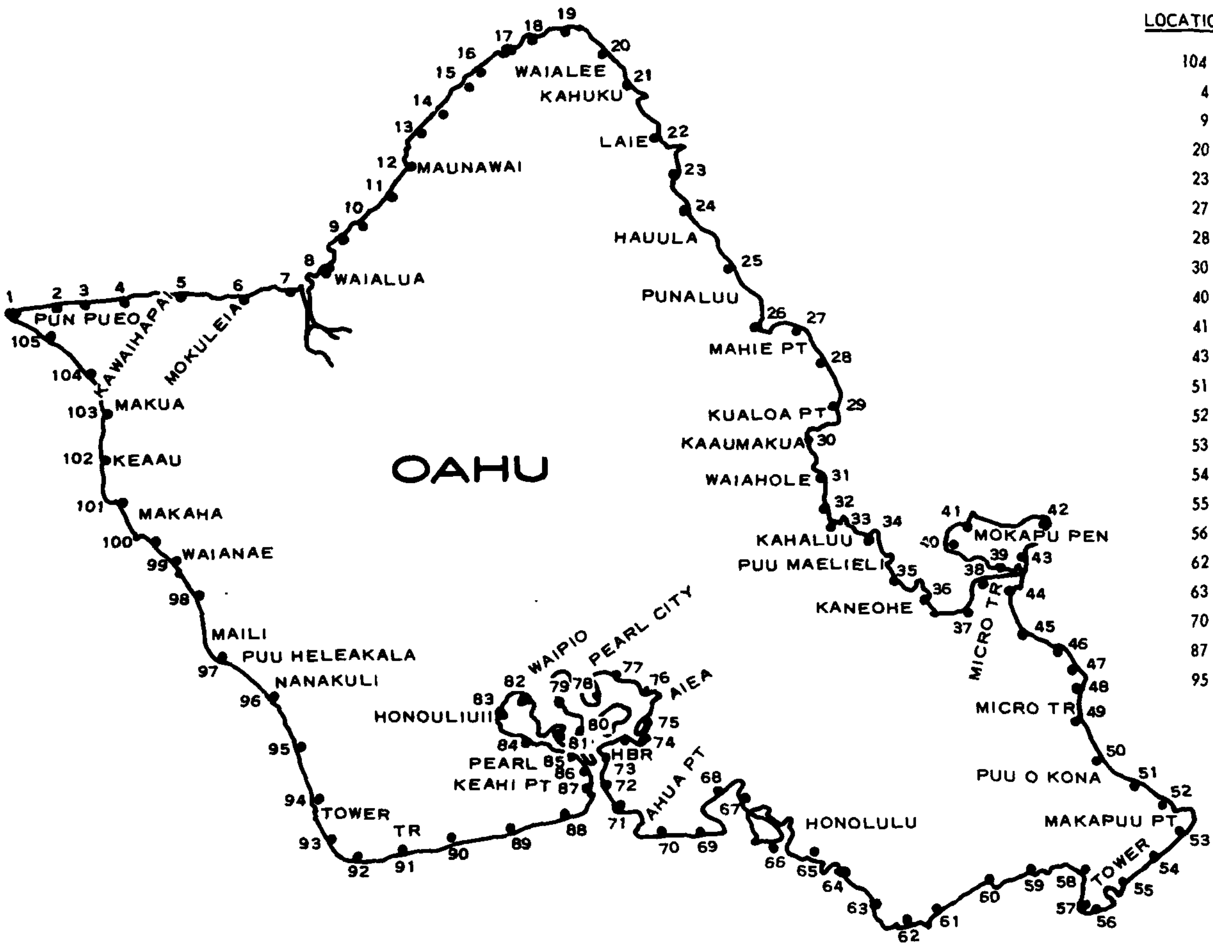


Figure 4. Tsunami hazard for Oregon and Washington (adapted from Houston and Garcia, 1978)



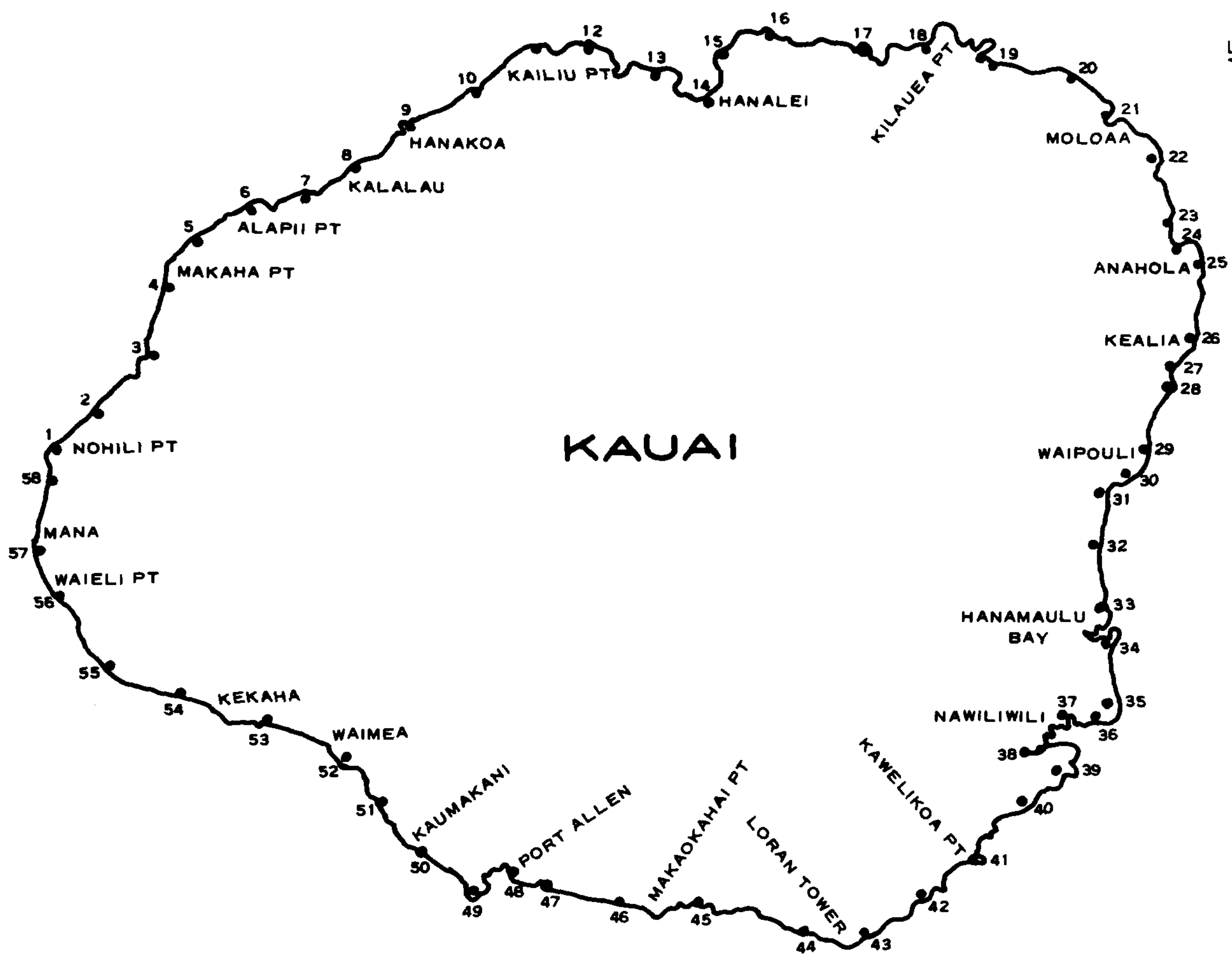
LOCATION NUMBER	ZONE
154 TO 4	3
4 TO 11	2
11 TO 13	3
13 TO 17	2
17 TO 19	3
19 TO 27	4
27 TO 31	3
31 TO 48	2
48 TO 61	3
61 TO 62	4
62 TO 64	5
64 TO 66	3
66 TO 69	4
69 TO 70	5
70 TO 77	4
77 TO 82	3
82 TO 110	4
110 TO 124	3
124 TO 127	4
127 TO 130	3
130 TO 134	4
134 TO 141	5
141 TO 154	4

Figure 5. Tsunami hazard map for Island of Hawaii (adapted from Houston et al., 1977)



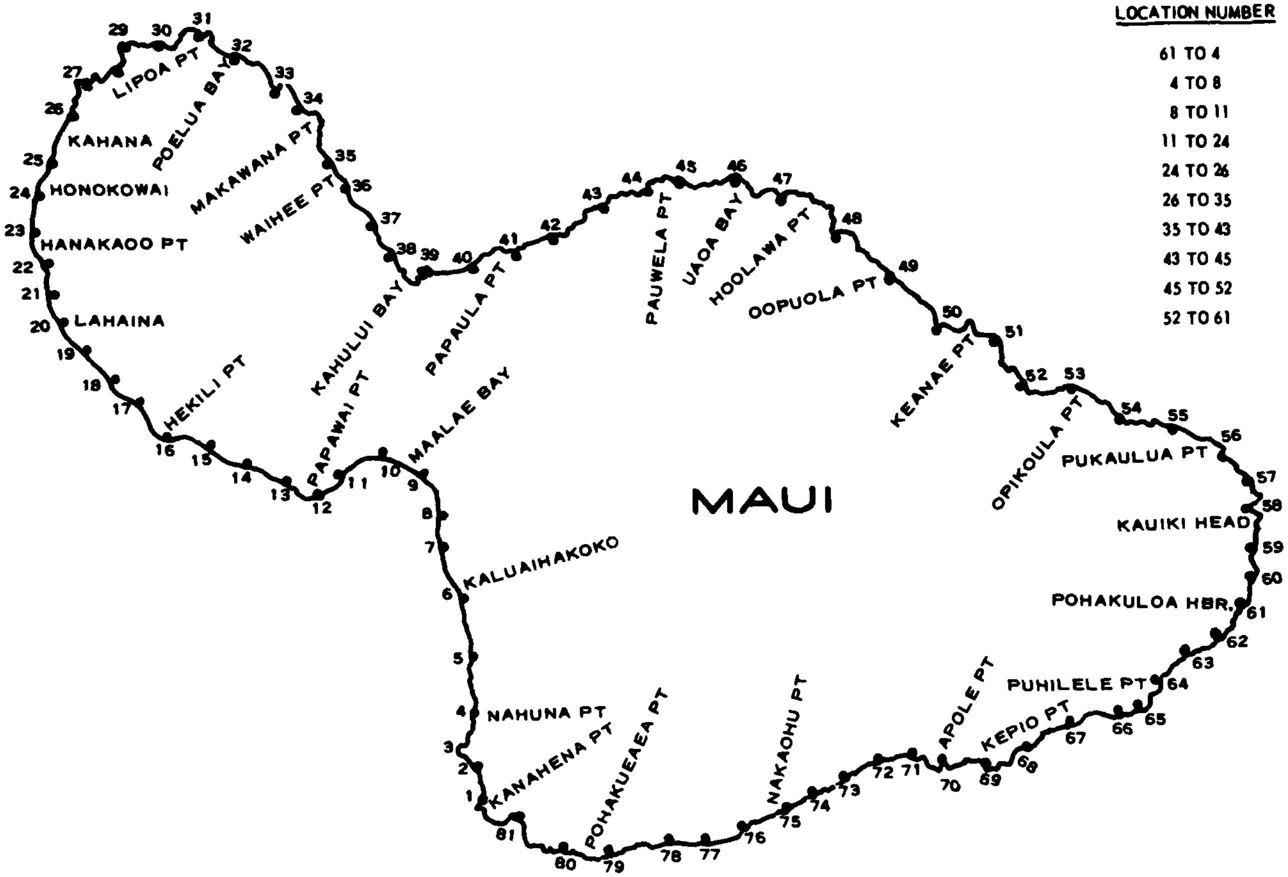
LOCATION NUMBER	ZONE
104 TO 4	4
4 TO 9	3
9 TO 20	4
20 TO 23	3
23 TO 27	2
27 TO 28	3
28 TO 30	2
30 TO 40	1
40 TO 41	2
41 TO 43	3
43 TO 51	2
51 TO 52	3
52 TO 53	4
53 TO 54	3
54 TO 55	4
55 TO 56	3
56 TO 62	2
62 TO 63	3
63 TO 70	2
70 TO 87	1
87 TO 95	2
95 TO 104	3

Figure 6. Tsunami hazard map for Oahu (adapted from Houston et al., 1977)



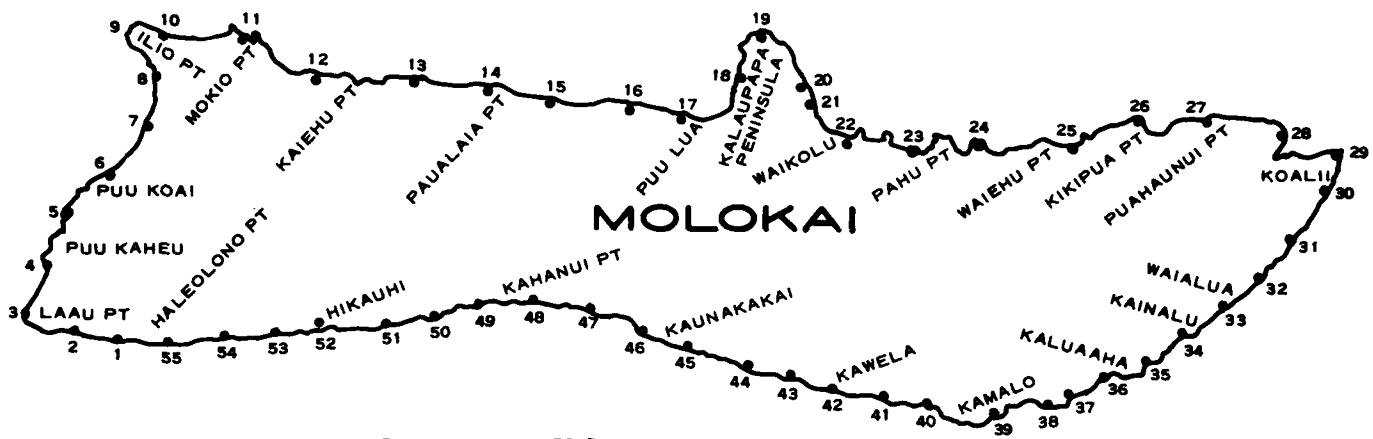
LOCATION NUMBER	ZONE
1 TO 4	5
4 TO 5	4
5 TO 10	3
10 TO 11	4
11 TO 13	5
13 TO 14	4
14 TO 15	3
15 TO 16	4
16 TO 17	3
17 TO 18	4
18 TO 19	5
19 TO 22	4
22 TO 24	3
24 TO 26	4
26 TO 27	3
27 TO 28	4
28 TO 34	3
34 TO 35	4
35 TO 37	3
37 TO 45	2
45 TO 48	3
48 TO 54	2
54 TO 57	3
57 TO 1	4

Figure 7. Tsunami hazard map for Kauai (adapted from Houston et al., 1977)



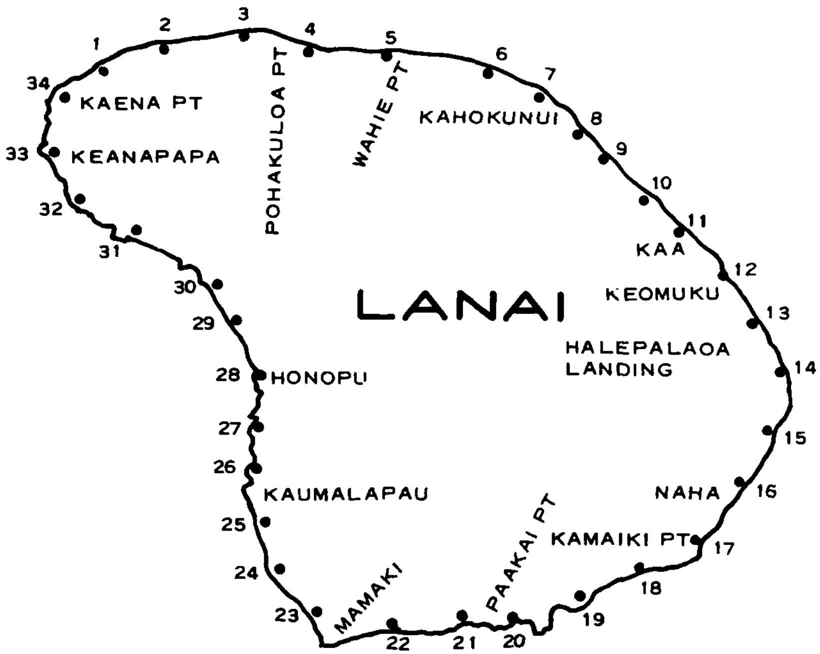
LOCATION NUMBER	ZONE
61 TO 4	3
4 TO 8	2
8 TO 11	3
11 TO 24	2
24 TO 26	3
26 TO 35	4
35 TO 43	3
43 TO 45	4
45 TO 52	3
52 TO 61	4

Figure 8. Tsunami hazard map for Maui (adapted from Houston et al., 1977)



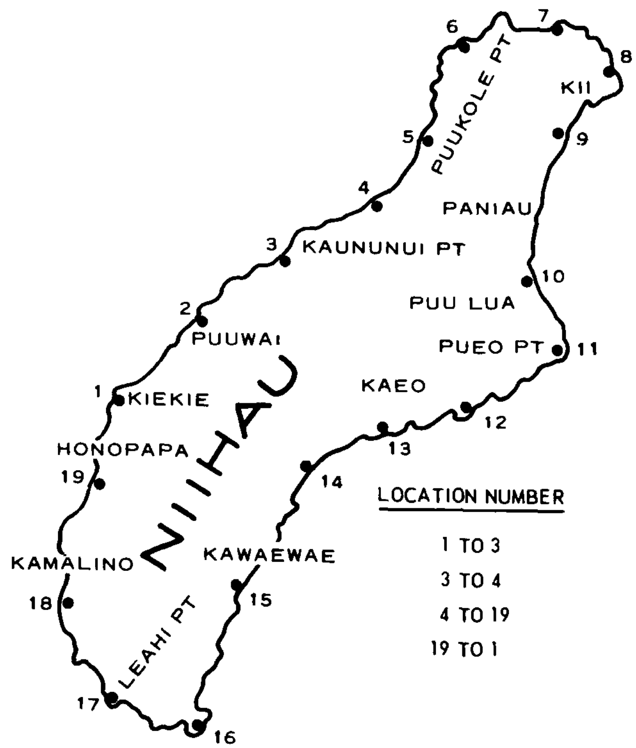
INTERVAL	ZONE
55 TO 2	3
3 TO 4	4
4 TO 5	3
5 TO 18	4
18 TO 21	3
21 TO 22	4
22 TO 26	5
26 TO 31	4
31 TO 32	3
32 TO 37	2
37 TO 52	1
52 TO 55	2

Figure 9. Tsunami hazard map for Molokai (adapted from Houston et al., 1977)



LOCATION NUMBER	ZONE
1 TO 6	3
6 TO 10	2
10 TO 14	3
14 TO 1	2

Figure 10. Tsunami hazard map for Lanai (adapted from Houston et al., 1977)



LOCATION NUMBER	ZONE
1 TO 3	3
3 TO 4	4
4 TO 19	3
19 TO 1	3

Figure 11. Tsunami hazard map of Niihau (adapted from Houston et al., 1977)

PART V: THE TSUNAMI WARNING SYSTEM

92. Although tsunamis propagate at velocities as great as 500 mph in the deep ocean, transoceanic distances are sufficiently large to allow several hours warning before arrival of a distantly generated tsunami. Thus, given prompt warning of the generation of a tsunami, there is sufficient time to reduce the potential hazard to life or mobile equipment (such as ships or automobiles) by evacuation procedures. The Tsunami Warning System (TWS) was founded in 1946 (following the Aleutian tsunami of 1 April 1946 that caused major damage and many casualties in the Hawaiian Islands) by the U. S. Coast and Geodetic Survey to provide warning in the event of the occurrence of a tsunami. The TWS is currently operated by the National Weather Service, which is under the direction of the National Oceanic and Atmospheric Administration of the U. S. Department of Commerce.

93. The TWS is a cooperative effort among nations bordering the Pacific Ocean to provide early warning of potentially disastrous tsunamis. Seismograph and tide data are collected and communicated to the Pacific Tsunami Warning Center (PTWC) at Ewa Beach, Hawaii, for analysis and evaluation. If warranted, tsunami watches or warnings are disseminated immediately by the Center to affected coastal populations (Spaeth, 1975).

94. The TWS begins functioning with the detection, by any participating seismic observatory, of an earthquake of sufficient size to trigger the alarm attached to the seismograph at that station. The alarm thresholds are set for each station so that the ground vibrations of the amplitude and duration associated with an earthquake of approximate magnitude of 6.5 or greater anywhere in the Pacific region will cause them to sound. This magnitude is below the threshold for issuing watch and warning messages. Personnel at the station immediately interpret their seismographs and send their readings to the PTWC. Upon receipt of a report from one of the participating seismic observatories or as a consequence of the triggering of their own seismic alarm, PTWC

personnel send messages requesting data to the observations in the system (Spaeth, 1975).

95. When sufficient data have been received for PTWC personnel to locate the earthquake and compute the magnitude, a decision is made as to further action. If the earthquake is strong enough to cause a tsunami and is located in an area where tsunami generation is possible, PTWC personnel will request participating tide stations located near the epicenter to monitor their gages for evidence of a tsunami. Watch bulletins are issued to the dissemination agencies for earthquakes of magnitude 7.5 or greater (7 or greater in the Aleutian Island region), alerting them to the possibility that a tsunami has been generated and providing data that can be relayed to the public so necessary preliminary precautions can be taken. A watch also may be disseminated by the PTWC upon issuance of warnings by regional warning centers. Since the regional systems use different criteria for their disseminations, a watch may at times be issued by the PTWC for earthquakes with magnitudes less than 7.5 (Spaeth, 1975).

96. When reports are received from tide stations, they are evaluated; if they show that a tsunami has been generated that poses a threat to the population in part or all of the Pacific, a warning is transmitted to the dissemination agencies for relay to the public. The dissemination agencies then implement predetermined plans to evacuate people from endangered areas. If the tide station reports indicate that either a negligible tsunami or no tsunami has been generated, PTWC issues a cancellation of its previously disseminated watch (Spaeth, 1975).

97. Because of the time spent in collecting seismic and tidal data, the warnings issued by the PTWC cannot protect areas against tsunamis generated in adjacent waters. For providing some measure of protection against local tsunamis in the first hour after generation, regional warning systems have been established in some areas. To function efficiently, these regional systems generally have data from a number of seismic and tide stations telemetered to a central headquarters. Nearby earthquakes are located, usually in 15 min or less, and a warning based on seismological evidence is released to the population of the area.

Since the warning is issued on the basis of seismic data alone, one may anticipate that warnings occasionally will be issued when tsunamis have not been generated. Since the warnings are issued only to a restricted area and confirmation of the existence or nonexistence of a tsunami is obtained rapidly, dislocations due to the higher level of protection are minimized. Among the most sophisticated of the regional systems are those of Japan and Alaska. The command center of the Alaska regional system is the Alaska Tsunami Warning Center in Palmer, Alaska (Spaeth, 1975).

98. Offices of the U. S. Army Corps of Engineers may obtain information about tsunami watches and warnings from the U. S. Army Engineer Division, Pacific Ocean; the U. S. Sixth Army, Presidio, California; or Civil Defense agencies of Hawaii, California, Oregon, Washington, or Alaska. These organizations receive all tsunami watch and warning messages from the PTWC. Tsunami arrival times in the Hawaiian Islands from tsunamis generated anywhere in the Pacific region can be estimated by using Figure 12. Tsunami arrival times in Alaska, Hawaii, or the west coast of the United States for tsunamis generated in the Aleutian Islands or eastern Alaska can be estimated by using Figures 13 and 14. The PTWC does not supply predictions of areas in possible danger. However, elevation predictions by Houston et al. (1977) for the Hawaiian Islands, Houston and Garcia (1974) for southern California, Garcia and Houston (1975) for Monterey and San Francisco Bays and Puget Sound, and Houston and Garcia (1978) for the rest of the west coast of the United States can be used to determine areas of possible danger. Personnel should evacuate these areas if there is sufficient time; ships should depart harbors for the safety of the open seas (tsunami elevations are very small in the open ocean). Vessels should avoid harbor entrances or other restricted channels where large currents may develop during tsunami activity. Vessels also should be moored at locations away from important facilities if large tsunami elevations are expected since the vessels may act as projectiles and cause destruction.

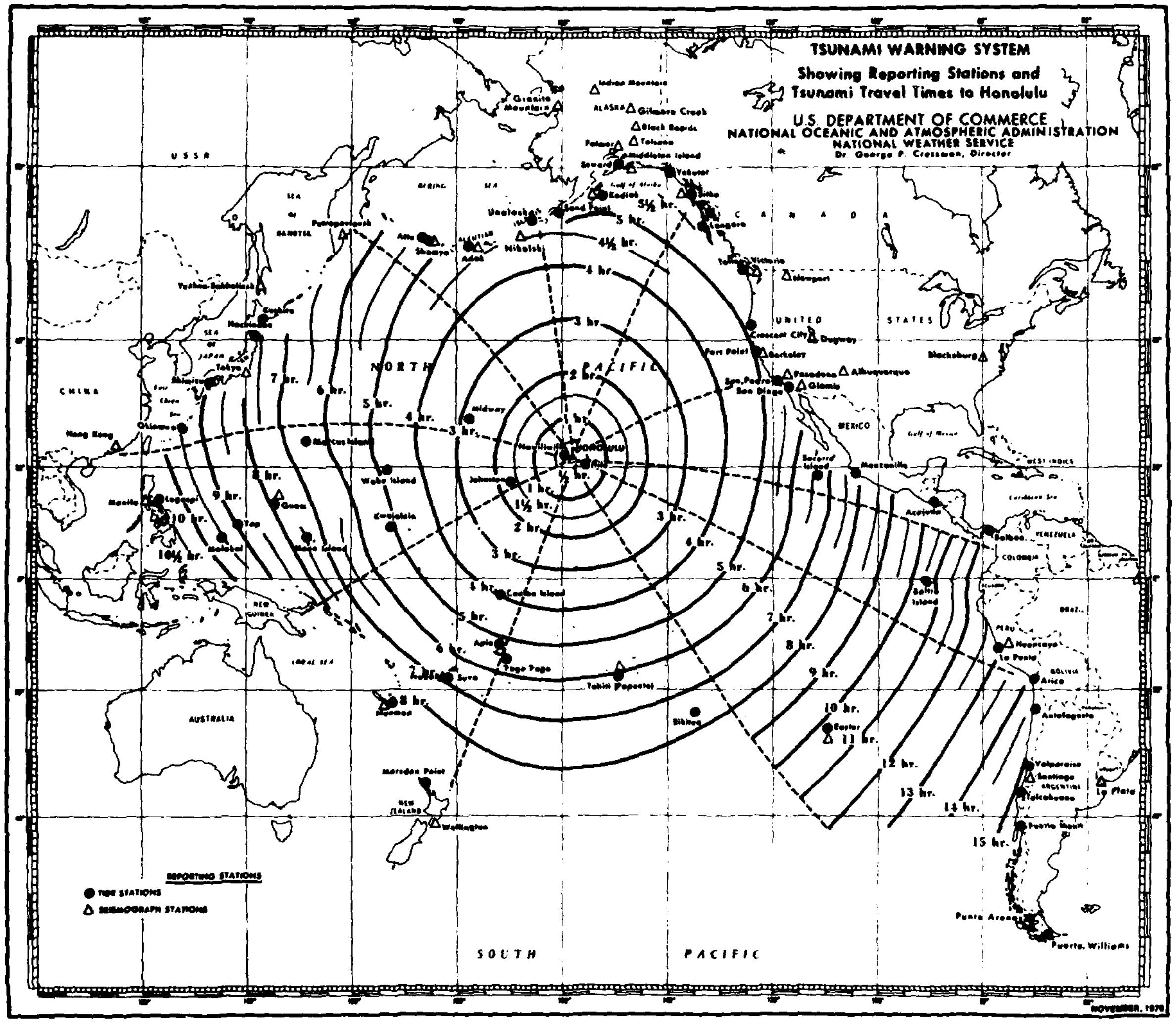


Figure 12. Arrival times to Honolulu (from Pararas-Carayannis, 1977)

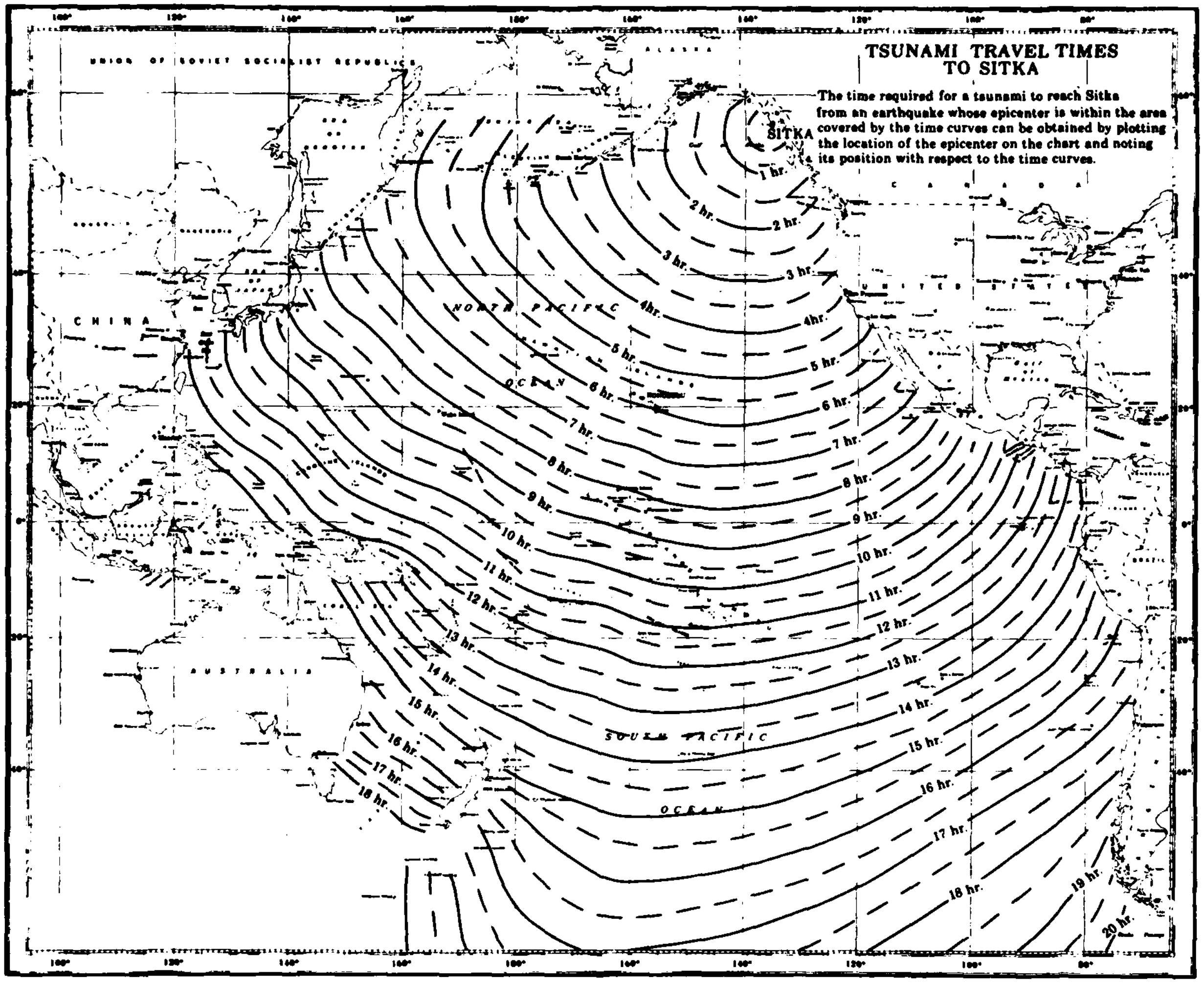


Figure 13. Chart of Pacific area with tsunami travel times to Sitka (from Cox and Pararas-Carayannis, 1976)

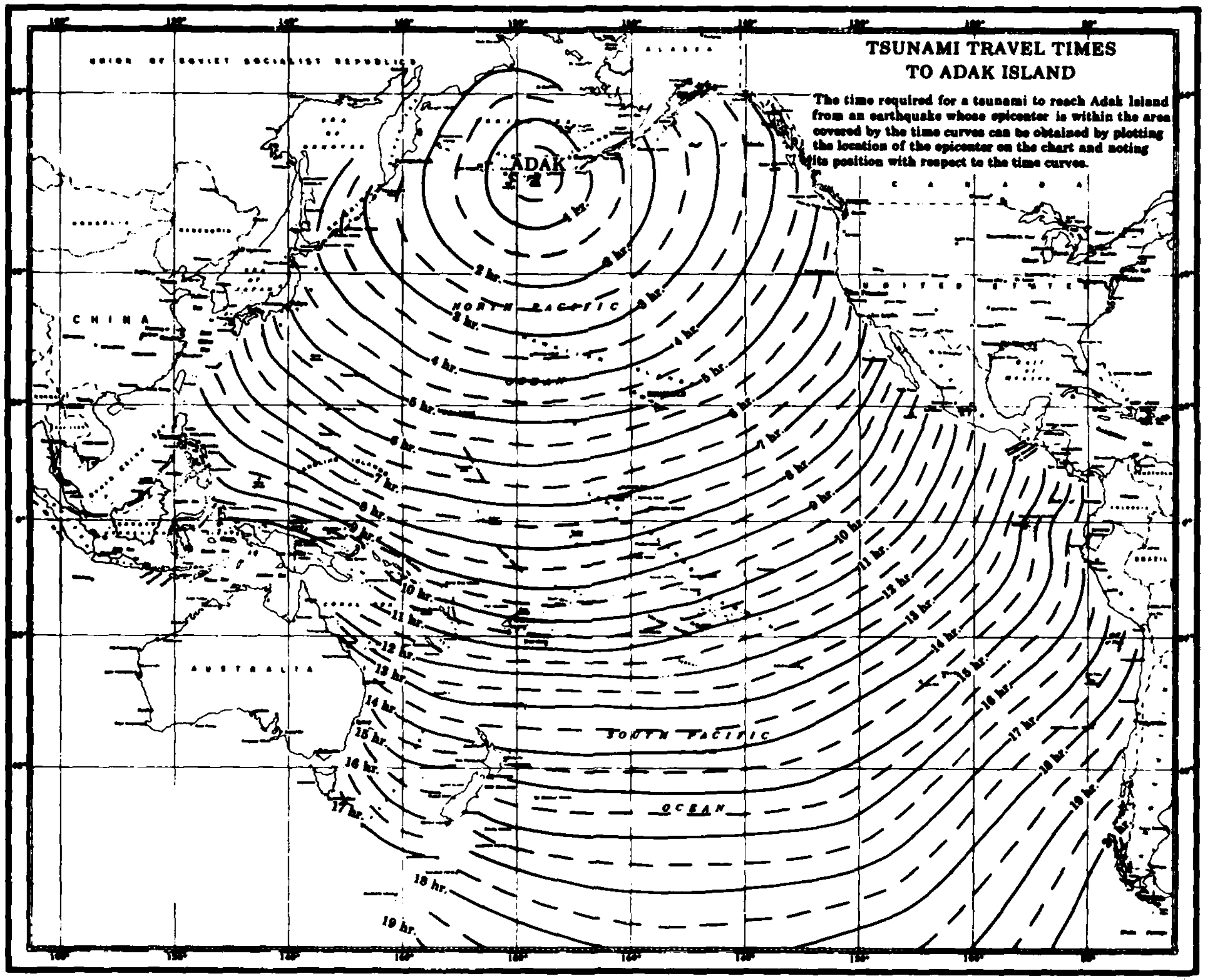


Figure 14. Chart of Pacific area with tsunami travel times to Adak Island (from Cox and Pararas-Carayannis, 1976)

PART VI: TSUNAMI STRUCTURAL DAMAGE

Introduction

99. Tsunamis can cause great damage to structures. The severe destruction in Hilo, Hawaii, during the 1946 and 1960 tsunamis (Shepard et al., 1950; Reese and Matlock, 1960); in many Alaskan cities during the 1964 tsunami (Wilson and Torum, 1968); and in Crescent City, California, during the 1964 tsunami (Wilson and Torum, 1968; Magoon, 1965) is well documented. The tremendous forces developed by tsunamis are illustrated by the destruction in Seward, Alaska, during the 1964 tsunami. Wilson and Torum reported that a 115-ton locomotive was overturned and transported 300 ft by this tsunami. A winch bolted down and welded to railroad irons set in 6 ft of concrete was torn loose by wave forces that sheared four pieces of railroad steel. The waves carried a 26-ton crane 500 ft inland and wrapped a 6-ton panel truck around a tree.

100. To calculate the force of tsunamis on a structure, it is necessary to know the speed and direction of flow, as well as the water level as a function of time. The nature and shape of the structure are, of course, of basic importance. Tsunamis usually appear as rapidly rising water levels. Wilson and Torum (1968) showed that for such a case the speed of the water across land is approximately

$$u = \frac{2\pi A}{Ts} \cos\left(\frac{2\pi t}{T}\right) \tag{3}$$

where

u = horizontal speed, ft/sec

A = tsunami amplitude, ft

T = tsunami period, sec

s = ground slope

t = time, sec

For small ground slopes, the tsunami speed becomes limited by the

propagation speed of a bore. This speed is given by the expression (Wilson and Torum, 1968)

$$u = C(gd)^{1/2} \tag{4}$$

where

C = dimensionless number varying between 1 and 2

g = acceleration of gravity, ft/sec^2

d = height of bore face above the local land elevation

For areas where two-dimensional flow effects are important, velocities (speed and direction) and water levels as a function of time can be determined using a land inundation numerical model (Houston and Butler, 1979). Maximum elevation predictions (Houston et al., 1977; Houston and Garcia, 1974 and 1978; Garcia and Houston, 1975) can be used as input to the numerical model. It can be assumed that the form of the tsunami maximum wave is approximately sinusoidal. A wave period can be selected from historic data of tsunami activity in the area.

Forces

101. Structures may be subjected to five major forces during a tsunami. These are bouyancy, hydrostatic, drag, surge, and impact forces. The buoyancy force is a vertical uplifting force acting on a structure due to the displacement of a volume of water equal to the volume of the structure that is submerged. The hydrostatic force is a horizontal force due to the weight and depth of water acting on the structure. The drag force results from the flow of water past an object. Surge forces are due to changes in either the direction or magnitude of flow and are important when the leading edge of a surge impinges on a structure. Impact forces are due to material carried by the flow of water impacting a structure (U. S. Army Engineer Division, Pacific Ocean, 1978).

102. The bouyant force is the weight of water displaced by a structure that is partially or totally submerged. The buoyant force on a structure is given by

$$F_B = \rho g V \tag{5}$$

where

F_B = bouyant force, lb

ρ = density of water, lb-sec^2/ft^4 (slugs/ft^3)

g = acceleration of gravity, ft/sec^2

V = submerged volume of the structure, ft^3

Figure 15 shows the relationship between bouyant force and floor area for a watertight building for various inundation heights. When the bouyant force exceeds the weight of the building, it will begin to float. An unanchored building with a floor area of 1500 sq ft with a

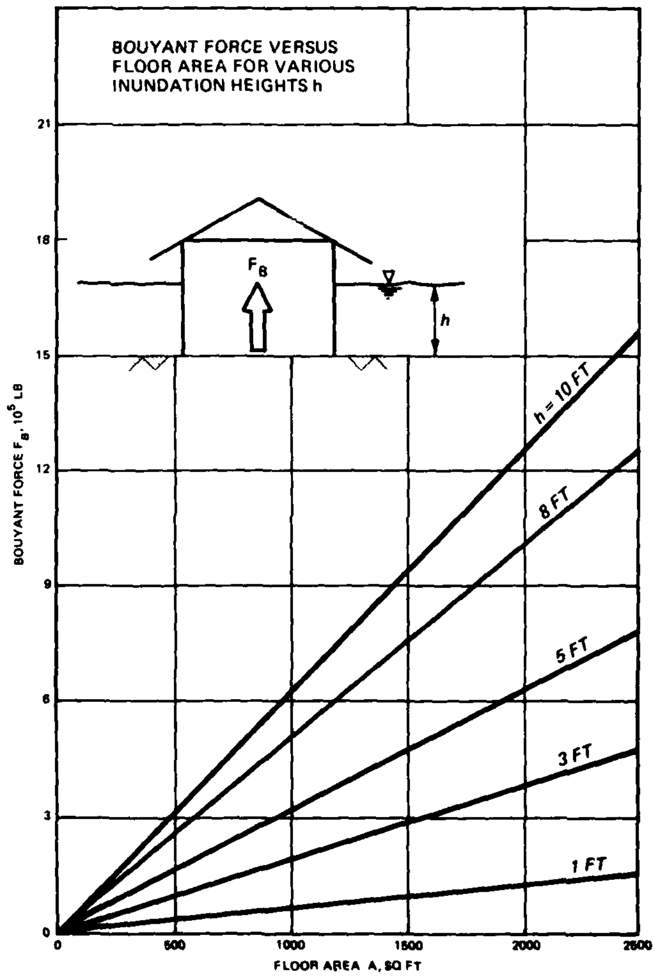


Figure 15. Bouyant force (from U. S. Army Engineer Division, Pacific Ocean, 1978)

"dead" load of 60 psf weights 90,000 lb. This does not include the additional weight of the building due to furnishing or other "live" loads. If the building remains watertight (no intrusion of water), it will begin to float at an inundation depth of only approximately 1 ft (U. S. Army Engineer Division, Pacific Ocean, 1978). Residential structures near the shoreline in Hawaii are often unanchored and are floated off their foundations during tsunamis. In many cases, this floating is fairly gentle, and sometimes houses can be returned to their foundations with little repair required. Shepard et al. (1950) reported a house at Kewela Bay, Oahu, being floated off its foundation during the 1946 tsunami and deposited in a cane field 200 ft inland, leaving breakfast cooking on the stove and dishes in place on shelves. Damage caused by buoyant forces during tsunamis is often the result of buildings being deposited on uneven ground or buildings breaking apart when they are lifted from their foundations as a result of weak structural integrity (Reese and Matlock, 1960).

103. Hydrostatic forces are significant when stagnant of slow-moving water piles up on one side of a wall or other barrier. The hydrostatic force is given by

$$F_H = \frac{1}{2} \rho g h A \tag{6}$$

where

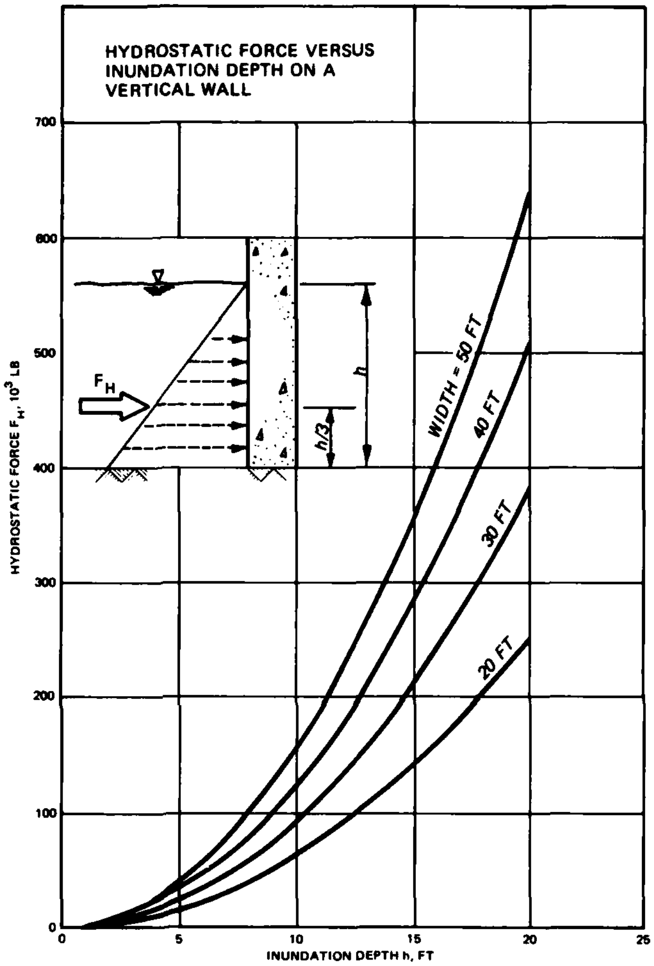
F_H = hydrostatic force, lb

h = depth of inundation, ft

A = projected area, ft^2 (depth of inundation h times the width of the structural member).

Since the hydrostatic pressure acting on a wall increases linearly with depth, the force vectors have a triangular distribution. The resultant force is located in Figure 16 at the centroid of the triangular area (a distance of h/3 from the base). Figure 16 shows the hydrostatic force on various widths of vertical wall for different inundation depths (U. S. Army Engineer Division, Pacific Ocean, 1978). An imbalance in water level of approximately 2 ft or greater can cause serious damage to a structure. The net hydrostatic force on a solid structural member

Figure 16. Hydrostatic force (from U. S. Army Engineer Division, Pacific Ocean, 1978)



is zero if the member is surrounded by a constant depth of water.

104. Flow of water past an object produces drag forces. The drag force is given by (U. S. Army Engineer Division, Pacific Ocean, 1978)

$$F_D = \frac{1}{2} \rho \ C_D \ Au^2 \qquad (7)$$

where

F_D = drag force, lb

C_D = coefficient of drag, dimensionless

Drag coefficients for flow as high Reynolds numbers are generally not

available, but Camfield (in preparation) used known drag coefficients to establish maximum coefficients for design pruposes. Coefficients generally vary from approximately 0.5 to 2.0. The drag coefficient for a vertical wall is equal to 2.0. Figure 17 shows the force due to drag on a 12-in.-diam pole.

105. Surge forces are important when the leading edge of a surge impinges on a structure. The surge force is given by

$$F_S = \rho\ C_S\ Au^2 \tag{8}$$

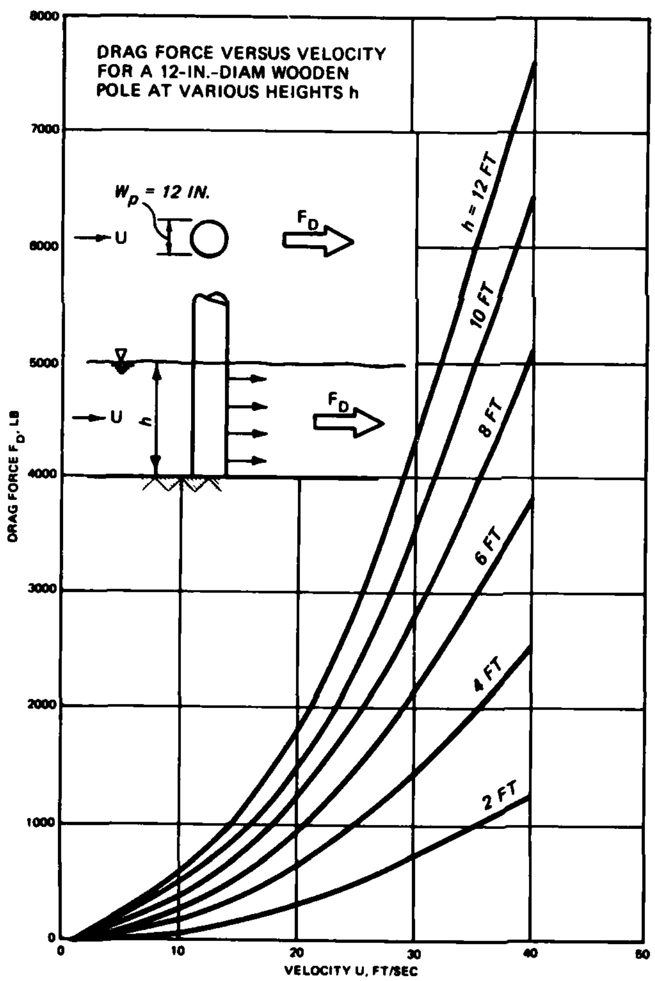


Figure 17. Drag force (from U. S. Army Engineer Division, Pacific Ocean, 1978)

where

F_S = surge forces, lb

C_S = surge coefficient, dimensionless

Cross (1967) relates C_S to the inclination of the water surface by the equation

$$C_S = (\tan \theta)^{1.2} + 1 \tag{9}$$

where θ is the inclination of the water surface of the surge. Therefore, $\tan \theta$ is equal to the rate of change of the height of the surge with the distance from the surge front. C_S equals 1 for a fairly flat surge face.

106. Impact forces are important when objects are floated by the tsunami and impact structures. These objects may include trees, boulders, automobiles, boats, and even buildings. Magoon (1965) indicated that substantial damage occurred in Crescent City, California, during the 1964 tsunami as a result of the collision of logs and lumber with structures. During the 1964 tsunami, boats in the harbor at Kodiak City, Alaska, were rammed into buildings along the waterfront. At Whittier, Alaska, small tanks at a petroleum tank farm that were picked up by the 1964 tsunami impacted buildings and larger tanks. The larger tanks would have been able to withstand the direct effects of the tsunami flow but were set into motion by the impact of the smaller tanks and ruptured. A resulting fire destroyed the tank farm (Camfield, in preparation).

107. The impact force caused by an object moving with the velocity of the tsunami flow is expressed as

$$F_I = m \frac{u_2 - u_1}{\Delta t} \tag{10}$$

where

F_I = impact force, lb

m = mass of the impact object, lb-sec^2/ft (slugs)

u_1 = speed of water flow (or speed of the object if it is moving at a lesser speed), ft/sec

u_2 = speed of object after impact with a structure (equal to zero if structure remains fixed after impact), ft/sec

Δt = time interval of impact, sec

108. The impact time may be very short if both the object and the structure are rigid. If either the object or structure deforms upon impact, the time interval may be significantly greater than for the collision of rigid objects. In the handbook published by the U. S. Army Engineer Division, Pacific Ocean (1978), the time interval of impact is estimated to usually vary between 0.01 to 1.0 sec. Figure 18 shows the relationship between velocity and impact force for a 1000-lb object and a 1.0-sec impact time.

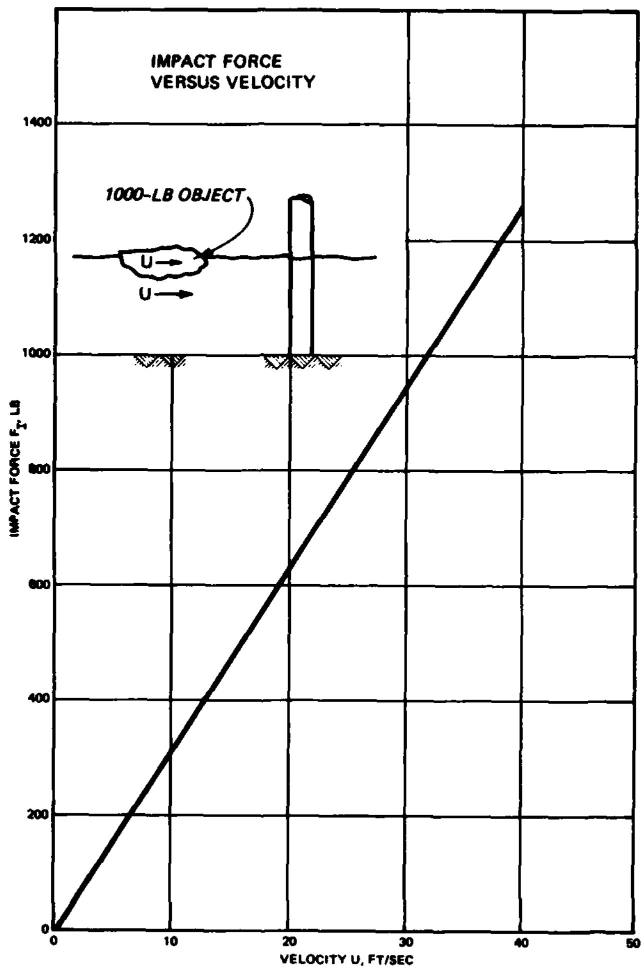


Figure 18. Impact force (from U. S. Army Engineer Division, Pacific Ocean, 1978)

PART VII: SEICHES

Introduction

109. Seiching of enclosed bodies of water often occurs following an earthquake. Reid (1914) reported that the power of an earthquake to cause seiching in lakes, ponds, and canals was first noticed at the time of the Lisbon earthquake of 1755. This earthquake excited seiching motions in small bodies of water throughout Europe as far away from Portugal as Finland. Reid attributed the seiching motion to a synchronism of period between the long-period surface waves produced by the earthquake and natural periods of oscillation of the bodies of water. The ability of earthquakes to produce seiching in bodies of water at tremendous distances from the epicenter also was noted during the Assam (eastern India) earthquake of 1950 that produced seiching in Norway and England (Kvale, 1955).

110. The 1964 Alaskan earthquake caused extensive seiche action in lakes, ponds, canals, rivers, waterways, and swimming pools all along the Louisiana coast and along the Texas coast as far west as Freeport. Although damage was gradually minor, it was very widespread along this part of the Gulf Coast of the United States. There were many reports of boats breaking their mooring lines and sinking or damaging piers, docks, and marine service facilities. A small U. S. Corps of Engineers boat was destroyed near Morgan City, Louisiana, when it was crushed between another boat and a tree on the bank as a result of a 4-ft seiche. Waves 1 to 5 ft in height were reported at many locations (Spaeth and Berkman, 1967). Donn (1964) and McGarr (1965) attributed these seiches reported on the Gulf Coast and seiches reported in two reservoirs in Arkansas to excitations induced by arrival of Love or Rayleigh waves. The amplitudes of these surface waves were very large on the Gulf Coast with almost all of the seismographs in this region used to measure surface waves driven off scale. The unusually large amplitudes are believed to be related to the great thickness of sediment in the Gulf Coast region. The amplitude of the seiching motion induced in a body of water is

dependent upon the amplitude of the long-period surface waves generated by the earthquake and the similarity between the period of the surface waves and the natural periods of oscillation of the body of water.

111. More dramatic seiching motion may be produced by a permanent vertical ground displacement occurring in an enclosed body of water. For example, on 17 August 1959 an earthquake with a magnitude of 7.1 occurred at Hebgen Lake in southern Montana. It has been estimated that the lake bottom dropped a distance of 10 ft at Hebgen Dam and up to 19 to 21 ft approximately 4 miles upstream (Wiegel and Camotim, 1962). This displacement of the bottom of the lake produced a seiching motion in the lake. At the dam, water alternately flowed over the dam and then flowed away leaving exposed the lake bottom near the dam. Water probably flowed over the dam four times with a maximum height of 4 ft or more. Each flow of water over the dam lasted from 10 to 20 min. The rush of water down the back of the dam produced some erosion, particularly at the base of the dam. A small powerhouse at the base of the dam was filled with debris carried by the water, and the spillway (which also may have been directly damaged by the earthquake) was damaged as a result of the erosion of earth beneath some concrete slabs (Steinbrugge and Clould, 1962).

Modeling of Seiches

112. One of the few attempts to model seiching activity in an enclosed body of water induced by earthquake activity was a hydraulic model investigation by Wiegel and Camotim (1962) of seiching in Hebgen Lake during the 17 August 1959 earthquake. This model was constructed of wood and was distorted, with the horizontal scale about 1:8000 and the vertical scale about 1:3500. The sidewalls of the model were vertical, and the bottom was either sloping uniformly or horizontal. The action of the earthquake was simulated by dropping the end of the model corresponding to the dam. The small scale of the model resulted in severe damping; thus, the duration and amplitudes of the seiching could not be accurately modeled.

113. Numerical models can be used to study seiching in enclosed bodies of water. Since the oscillations have a long period, long waves govern the motion. Seiching induced by horizontal shaking of an enclosed body of water can be studied using a numerical model developed by Lee and Hwang (1978). The enclosed body of water must have a constant depth. Seiching produced by a permanent vertical ground displacement can be studied using any of the tsunami numerical models discussed in Part III. Since reservoirs usually have very complicated shapes, a finite element model, such as that developed by Kawahara et al. (1978), would be advantageous since the finite element method uses very flexible grids that allow complicated shapes to be well represented. Finite element models, such as those developed by Desai and Abel (1972) and Patil (1970), that solve free oscillation problems also can be used for seiching problems. These models require little computational time. The model of Desai and Abel has been used to study seiching induced by atmospheric phenomena in Lake Erie by Raney et al. (1974). An arbitrary vertical displacement can be separated into free modes of oscillation and elevations at later times throughout the reservoir region determined using an approach described by Wilson (1975).

PART VIII: LANDSLIDE-INDUCED WATER WAVES

Introduction

114. Landslide-induced water waves can occur in coastal regions, enclosed bodies of water such as lakes or reservoirs, and rivers. These waves are a potential problem where steep inclines border water. Landslides can be generated by earthquakes, water saturation of soil, and other effects. Slides down marine slopes, known as subaqueous slides, also can generate destructive waves.

115. Landslide-induced water waves have produced great destruction in the past. In 1792, a landslide of approximately 700 million cubic yards of rock and soil fell into Shimabara Bay, Japan. Waves as large as 33 ft produced destruction along 50 miles of the shores of the bay. Trees with diameters as great as 9 ft were snapped by the tremendous forces. More than 15,000 people were killed, most of them by the waves. In 1756, a landslide in Langfjord, Norway, produced waves with heights as great as 130 ft. Effects of the waves were noticed at distances of 25 miles, and 32 people were killed. A series of landslides in Loen Lake and Tafjord, Norway, from 1905 through 1934 produced waves that killed over 178 people (Miller, 1960).

116. One of the great disasters of modern times occurred in 1963 when a rock slide fell into a reservoir in the Vaiont Valley, Italy. Approximately 310 million cubic yards of rock slid into the reservoir and filled the first 1.5 miles of the reservoir. The slide displaced a tremendous quantity of water that flowed over the dam. Although the dam itself was essentially undamaged, the water that flowed over the dam formed a flood wave that destroyed the city of Langarome in the Piave Valley and killed 3000 people. A large water wave generated by the landslide also traveled upstream and damaged a number of buildings in two villages (Wiegel, 1976).

117. Several large landslides have generated significant waves in the United States. In 1905, a hanging glacier fell into Disenchantment Bay, Alaska, producing waves with heights as great as 115 ft. Several

slides between 1944 and 1953 fell into Franklin D. Roosevelt Lake in the Columbia River valley and produced waves with heights as great as 65 ft. The highest wave that has been documented was generated in 1958 by a landslide that was triggered by an earthquake and slid into Lituya Bay, Alaska. The landslide generated a wave that reached an elevation of 1740 ft on the opposite side of the bay and stripped the hillside of trees that were 50 ft high. Similar slides generated waves in Lituya Bay in 1853, 1874, and 1936 (Miller, 1960).

118. Subaqueous landslides triggered by the 1964 Alaskan tsunami caused widespread damage in Alaska. The city of Whittier, Alaska, was completely destroyed by waves as great as 104 ft in elevation generated by a submarine slide. Subaqueous slides produced runup elevations as great as 220 ft at Port Valdez, Alaska (Wilson and Torum, 1964).

119. In addition to generating large waves, landslides can cause serious problems by damming rivers and thus creating natural reservoirs. Once these reservoirs fill from the riverine flow, the dam is overtopped and a rapid washout of the natural dam can cause disasterous flooding downstream. Landslides into rivers occur quite often. For example, in 1970, a slide of about 2 million cubic yards of material moved downward about 3200 ft and then slid across the South Fork of the Smith River in California. A few weeks later, a slide of several hundred thousand cubic yards moved 1700 ft in a few seconds down to the Eel River in California and then continued 650 ft across the river channel (Wiegel, 1976). In 1927, a landslide at Gros Ventre, Wyoming, blocked a river and created a natural dam. Once the reservoir filled, the dam was overtopped and washed out rapidly, resulting in a disasterous flood that killed several people in Kelly, Wyoming. During the earthquake at Hebgen Lake, Montana, in 1959, a rockslide in the Madison Canyon downstream from the Hebgen Lake Dam blocked water that overflowed the dam during seiching of Hebgen Lake. Remembering the disaster created by the Gros Ventre landslide of 1927, the U. S. Army Corps of Engineers reduced the height of the rockslide at Madison Canyon and thus the likelihood of a serious downstream flooding that might occur if a reservoir formed behind the rockslide (Steinbrugge and Cloud, 1962).

Modeling of Landslides

Hydraulic models

120. In addition to general hydraulic model studies of the dropping of rectangular boxes (representing rockslides) into bodies of water by Wiegel et al. (1970), hydraulic model tests of landslides have been performed using scale models of actual reservoirs. For example, a study of possible high-speed rockslides into the reservoir created by the Mica Dam on the Columbia River was performed using a hydraulic model (Western Canada Hydraulic Laboratories Ltd., 1970). In addition to investigating the waves generated by landslides, this study investigated the erosion resistance of the dam during overtopping by these waves. A study of landslides into Lake Koocanusa, Montana, was performed by Davidson and Whalin (1974) using a 1:120 undistorted scale model of Libby Dam and Lake Koocanusa. The magnitude of wave heights, runup along the sides of the lake, and overtopping of the dam were determined for four potential landslides. Bags of iron ore and lead were mixed to reproduce the correct rock density, and the bags were placed on an inclined plane slope to obtain the correct elevation-volume-shape relationship. A major advantage of using hydraulic models for modeling landslides is that the large amplitude and short-period waves generated by the landslide are accurately modeled in an undistorted and large-scale model.

Numerical models

121. The relatively short wavelengths and large amplitudes of landslide-induced water waves make it difficult to model them numerically. Noda (1970) developed a one-dimensional numerical solution of an integral equation that described landslide-induced water waves generated by simple falling bodies. Koutitas (1977) developed a one-dimensional finite element model and applied it to the case of a long narrow channel where a lateral landslide caused a sudden change of its cross section. Koutitas and Xanthopoulos (1978) developed a two-dimensional finite element numerical model that solved nonlinear long-wave equations. Raney and Butler (1975 and 1977) developed a two-dimensional finite difference model that solved nonlinear long-wave equations. They made

comparisons between numerical model calculations and the hydraulic model tests conducted by Davidson and Whalin (1974). Reasonable agreement was found between the leading (largest) wave calculated by the numerical model and measurements of Davidson and Whalin. Raney and Butler (1977) concluded that for small landslide velocities, the viscous drag and pressure drag contributions to wave heights and velocities are small, and that the physical displacement of the water by the landslide is the predominant factor. The water waves produced by the slide for this condition will generally propagate faster than the motion of the slide; thus, the initial phase of the landslide as it enters the water must be accurately defined if good numerical results are to be obtained. However, for large landslide velocities, significant contributions to wave heights and propagation velocities are produced by the viscous drag and pressure drag. Viscous and pressure drag contributions to wave heights and velocities are focused in the direction of the landslide to a greater degree than the contributions from the physical displacement of water. If the landslide is moving faster than the normal propagation velocity of the water waves it produces, the entire path and time history of the landslide becomes important in obtaining an accurate prediction of the first wave crest.

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APPENDIX A: NOTATION

A	Tsunami amplitude, ft; projected area, ft^2
C	Dimensionless number
C_D	Coefficient of drag, dimensionless
C_S	Surge coefficient, dimensionless
d	Height of bore face above local land elevation
D	Number of years
F	Frequency per year of tsunami occurrence
F_B	Buoyant force, lb
F_D	Drage force, lb
F_H	Hydrostatic force, lb
F_I	Impact force, lb
F_S	Surge force, lb
g	Acceleration of gravity, ft/sec^2
h	Depth of inundation, ft
H_a	Tsunami wave height in the direction of the major axis of a source of length a
H_b	Tsunami wave height in the direction of the minor axis of a source of length b
H_{ave}	Average runup over a coast, m
i	Tsunami intensity
i_1	Tsunami intensity for particular tsunami
i_2	Tsunami intensity for particular tsunami
m	Mass of impact object, $lb\text{-}sec^2/ft$ (slugs)
P	Probability
s	Ground slope

t	Time, sec
T	Tsunami period, sec
Δt	Time interval of impact, sec
u	Horizontal speed, ft/sec
u_1	Speed of water flow, ft/sec
u_2	Speed of object after impact with a structure, ft/sec
V	Submerged volume of the structure, ft^3
θ	Inclination of the water surface of a surge
ρ	Density of water, $lb\text{-}sec^2/ft^4$ ($slugs/ft^3$)

In accordance with letter from DAEN-RDC, DAEN-ASI dated 22 July 1977, Subject: Facsimile Catalog Cards for Laboratory Technical Publications, a facsimile catalog card in Library of Congress MARC format is reproduced below.

Houston, James R

State-of-the-art for assessing earthquake hazards in the United States; Report 15: Tsunamis, seiches, and landslide-induced water waves / by James R. Houston. Vicksburg, Miss. : U. S. Waterways Experiment Station ; Springfield, Va. : available from National Technical Information Service, 1979.

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